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River-aquifer interaction in the Middle Yobe River Basin, North East Nigeria

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This thesis is submitted in fulfilment of the requirement for the degree of Doctor of Philosophy

DEDICATION

This work is dedicated to Ruqayyatu Alkali, my younger sister,

who has never tested the sweet love of our late father,

Alkali Abubakar Usman.

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ABSTRACT

Development of the shallow alluvial groundwater of the Hadejia-Jama'are-Yobe River valley flood plain (northern Nigeria) has been taking place with increasing intensity over the past decade. However little has previously been known about the nature of the Yobe River-aquifer interaction, including the recharge mechanisms. This thesis reports on a study of the river-aquifer interaction in the middle section of the basin, centred on a field site near Gashua, Yobe State.

Detailed field studies were undertaken over a period of 14-months, which involved geoelectical sounding, drilling and water level monitoring. The results of the field study show that the Yobe basin is underlain by a sand and gravel aquifer, which is covered by an average of 1-3m of clay.

The Yobe River is in hydraulic continuity with the adjacent alluvial aquifer and variations in aquifer storativity have been recognised as an important factor in understanding the hydraulic behaviour of the Yobe River-alluvial groundwater system. A confined-unconfined groundwater regime exists within the present site and is an inherent characteristic of the alluvial groundwater system. The recognition of this state enables a conceptual flow model of the system to be developed. A multi-layer, spatially distributed model is proposed, in which transitions between confined and unconfined conditions can be realistically represented. A numerical model needs to be designed with these concepts in mind in order to simulate the system.

The research techniques employed in the study are appropriate for the assessment of the Yobe basin system where detailed data is currently not available. The combination of detailed geophysical survey, water level monitoring and conceptual modelling has led to a good understanding of the Yobe River-alluvial aquifer interaction. It is for this reason that the techniques employed in this study can be adapted for investigating the remainder of the basin downstream of the present site.

CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF PLATES

CHAPTER ONE

INTRODUCTION

1.1 Background

The general reduction in rainfall over the last two decades (1970-1990) accompanied by reduced river flows, has had a tremendous effect on the socio-economic and agricultural development of northern Nigeria. It has impeded rainfed agriculture and has caused general scarcity of water for both irrigation and domestic water supplies.

Recognising the severity of the potential problems, Federal and State agencies, as well as international lenders such as the World Bank, started seeking all possible means of developing and reducing uneconomic use of the limited resources to enhance food production and to provide reliable water supplies for domestic purposes. However many of the large irrigation schemes initiated did not produce the desired results (Carter et al,1983; Adams 1991; Carter, 1992). As a result emphasis is now placed on developing alluvial aquifers for both small scale irrigation and rural water supply.

The expanding need for water for small scale irrigation and rural water supply in northern Nigeria has accelerated development of alluvial groundwater. This form of development is rapid and extensive, and simple drilling techniques are used to drill several thousands of tubewells. For example Kano Agricultural and Rural Development Authority (KNARDA) since its inception (1983-1993) has drilled more than seven thousand tubewells, while Sokoto Agricultural Development programme (ADP) has established a target of sinking 1,800 tubewells per year (Kimmage, 1991). The schemes have increased both the irrigated land and crop output, and have given individual farmers the initiative to develop the alluvium privately.

Although the present approach is encouraging, there is every likelihood that problems will arise in the future as a result of the way shallow groundwater is being developed and managed. This is because the expanding developments of the alluvial aquifers have not been accompanied by commensurate scientific research. Most of the developments are done haphazardly and on an ad hoc basis without making an assessment of the aquifer properties and potentials, and setting up management strategies for both pumps and tubewell installations, as well as efficient and proper way of using water in the fields.

The future problems associated with this kind of development could be enormous. Private ownership of wells and pumps by farmers may lead to uncontrolled development and over exploitation, which may lead to increased pumping costs, reduced yields, and even complete failure of wells. There may also be the tendency for the farmers to over irrigate which may lead to water quality deterioration, waterlogging and salinisation of the irrigable land, and this will result in the land being abandoned or costly measure being necessary to reclaim it.

These kinds of problems have occurred in many arid and semi-arid regions where similar schemes were adopted. For example, the state of water wells in parts of Pakistan deteriorated (Bakiewicz et al,1985); while water tables in India are rising in some places due to inefficient use of water for irrigation, and are falling in other parts due to excessive pumping (Erez and de Ridder, 1977; Kavalanaker et al,1992).

Furthermore, inadequate information on the aquifers can lead to serious economic loss as a result of either poor well location and installation, or the complete absence of the water source. For example in a study (Kimmage, 1991), it was found in some villages along the Hadejia-Nguru floodplain that up to 70% of their tubewells are out of use.

Superimposed upon the foregoing are the effects of existing and proposed dams on the river flow regime and the consequent bearing on the socio-economic development of the downstream water users. For example, the completion of the Tiga dam in 1974 partially exposed the latent conflict existing between the upstream and downstream farmers, and is seemingly becoming almost an intractible problem with an ongoing struggle for the control of the shared resources pitting downstream users against upstream consumers. Downstream farmers often claim that the change of the runoff regime, thought to be presenting itself in reduced flooding, have affected their livelihoods by reducing their annual production of flood rice and recession crops. It is also likely that similar complaints will emerge in the near future when the popular, recently introduced, irrigation tubewells go dry or become depleted, which may occur as a result of reduced surface water flow, or as a consequence of inadequate development and management of the alluvial aquifer system. Groundwater development should, ideally, be undertaken only after careful planning which includes an analysis of the probable influence of the development on the surrounding area. Carefully planned groundwater development can result in more efficient and beneficial utilisation of limited water resources.

It is clear that the future of the river-aquifer system, subjected to both natural and man made interferences, and the entire small scale irrigation programme in the region will depend on the management strategies adopted both at the planning and operation stages.

Nevertheless, the entire concept of groundwater resource management-from the standpoint of both quantity and quality- presupposes the availability of an adequate data base upon which to build a management programme and administrative institutions. Regrettably, however, there is as yet no comprehensive, physical and management information on the Yobe River-alluvial aquifer system and such inadequacy engenders severe uncertainties regarding the range of decisions which management of this system must confront. Doubtless, therefore, the establishment of a comprehensive information base must be the first step in understanding the river-aquifer system which is subject to both natural (climatic) changes and artificial (dams and irrigation) interferences. Such information includes among others, a complete physical description and properties of the aquifers, quantitative data on the movement of both surface and groundwater, and most importantly understanding of the nature of the Yobe River-alluvial aquifer interaction.

Determination of river-aquifer interaction is a basic pre-requisite for efficient groundwater management since it must form one of the main inputs to any alluvial aquifer management model. It is particularly critical in northern Nigeria where massive and concentrated alluvial groundwater development is taking place, and yet no hard information exists on river-aquifer interaction.

It is apparent that a complete understanding of the Yobe River-alluvial aquifer interaction will enable us to design an optimal development scheme that neither overestimates the resources and creates difficulties during drought periods, nor underestimates the resources by neglecting the very high amount of recharge that could result from an important flood and be lost if the system has insufficient storage.

The present study was undertaken to provide a basis for improved management practices, to assist in appraising the abstraction potentials, and to define the river-aquifer system for safe and useful development of the alluvial groundwater.

1.2 Study aim and objectives

The aim of the study was to determine the nature of the Yobe River-alluvial aquifer interaction and so to develop a conceptual model of the system.

The study also aimed to achieve the following specific objectives:

1) to delineate the areal extent and disposition of the valley alluvium, and so to determine the sedimentary stratigraphy;

2) to identify the mechanisms of recharge to the alluvial groundwater;

3) to identify effects of confining layers and barriers in impeding recharge to, and groundwater movement in, the alluvial aquifer;

4) to evaluate the hydrogeologic properties of the aquifers; and

5) to examine the suitability of the geoelectric sounding technique for exploring the Yobe valley alluvium.

1.3 Study scope

The field study was deliberately confined only to a 4km length of the Yobe River valley, and extended over a period of one year. The results reflect only that section of the river valley and the condition of the river-alluvial aquifer system between May 1992 and June 1993.

The limitation of data was a constraint on the study. A multi-dimensional mathematical model could not be designed to represent the river-aquifer system, simply because there were not sufficient data. Nonetheless, a conceptual model of the system has been developed based on the available data. This model would serve as the framework on which to hang details and assist in collecting pertinent data for use in developing a mathematical model of the river-aquifer system.

1.4 Thesis structure

This thesis is divided into seven chapters. Chapter two gives a brief description of the study area (the Yobe basin). Chapter three covers a review of the water resources studies conducted previously in the Yobe basin. Chapter four contains the procedures and materials used in the present study. The geologic framework of the Yobe Valley alluvium is defined in Chapter five. The hydraulics of the system is presented in Chapter six. The conclusions and recommendations drawn from the study are presented in chapter seven.

CHAPTER TWO

THE STUDY AREA

This chapter gives brief descriptions of the physical settings and other features of the study area.

2.1 Location

The study area is the Hadejia-Jama'are-Yobe River basin. The river basin has an elongate shape west to east and covers about 135,000 km^2 (IWACO, 1985). The extensive inland plain covering the Hadejia-Jama'are-Yobe basin is part of the physiographic unit in Northern Nigeria referred to as the Chad basin (figure 2.1).

The study site is a transect across the floodplain of the middle section of the Hadejia-Jama'are-Yobe basin. It covers approximately 12 km^2 along a 4km length of the Yobe River immediately upstream of Gashua (figures 2.1 and 4.2).

2.2 Climate

The basin has a tropical climate. The climatic regime consists of a single dry season, October to April, followed by a shorter wet season.

Rain occurs normally in the months of June through September. Much of the rain falls as thunderstorms lasting frequently less than a few hours; periods of prolonged rainfall over several days are rare. Rainfall records show large fluctuations from year to year and from place to place. A 33 year record (1960-1992) at Nguru shows that the annual rainfall varied from more than 630mm in 1960 to less than 235mm in 1983. The spatial variation of rainfall in the basin is 1000mm to 1400mm on the south-west margin of the basin, decreasing to under 400mm on the NE margin around Lake Chad (Figure 2.2).

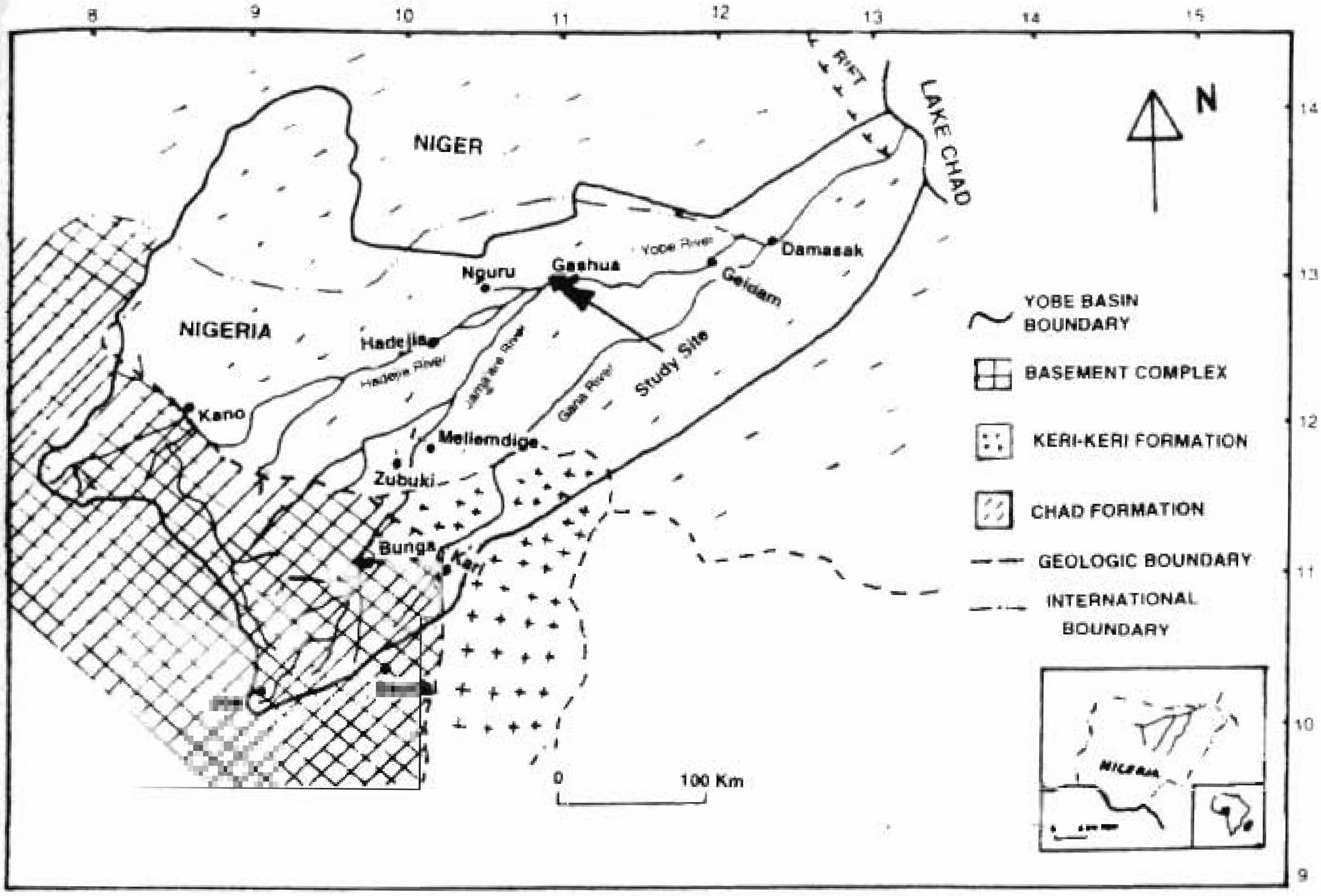


Figure 2.1. Location of the study area and major physical features

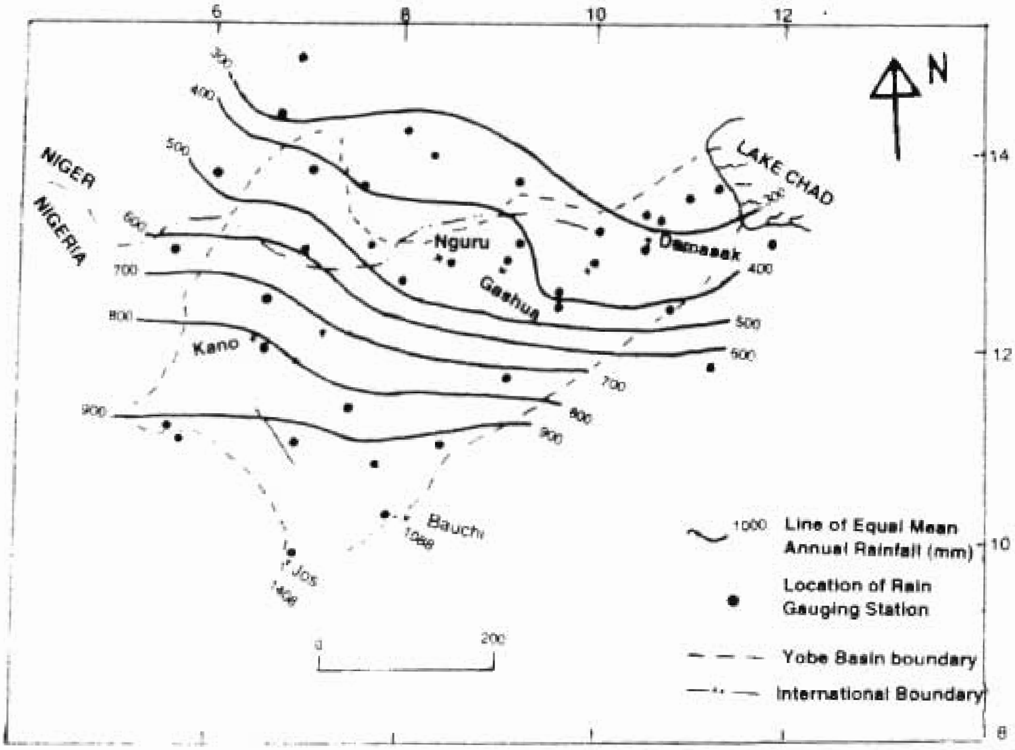


Figure 2.2. Spatial variation of mean annual rainfall over the Yobe Basin (After IWACO, 1985)

The dry season is hot, dry and windy. The combined effects of high radiation, high temperatures and frequent winds result in high evaporation rates. The annual class A pan evaporation rates recorded at Nguru is 3856mm/a. The corresponding evapotranspiration rates calculated using Penman method is 2348mm/a (Schultz, 1976).

2.3 Topography/Physiography

The topography of the lower part of the basin is characterised by a gently undulating plain with several shallow depressions. The land surface slopes gently from the high areas in the SW towards the NE at a gradient of less than 1m/km (IWACO, 1985). Locally the transverse slope is toward the rivers, and gently sloping upland which hardly exceeds 340m above sea level form the boundaries of the flooplain.

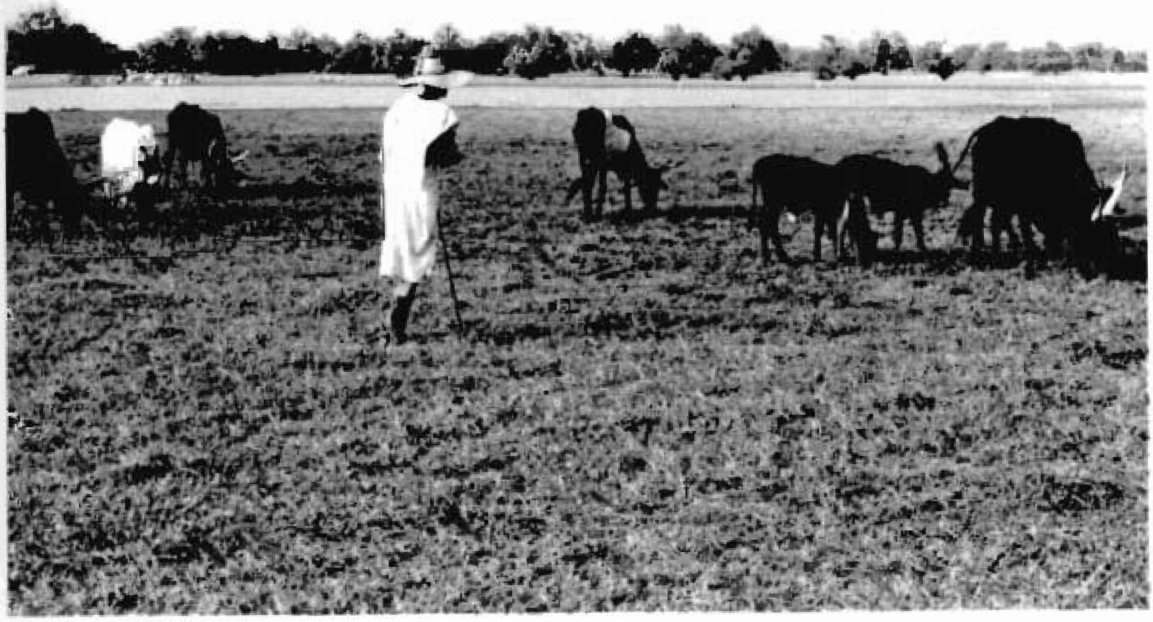
The modern floodplains are usually bounded by aeolian deposits, and themselves consist of

1) old channels, oxbow lakes and spillplains. These units are inundated even in relatively dry years; and

2) ridges (levees and other raised landforms) which are inundated only during above average floods.

The soils within the floodplain are clays and sandy/silty clays which are cracked especially at depression bottoms. The floodplain landuse is primarily agricultural; the fertile bottomland produces abundant yields of flood rice and recession crops, and rice stubbles and straws provide excellent pasture for cattle. In addition to farming, the local industries include fishing and sand mining. Plate 2.1 shows examples of agricultural practices within the Yobe valley.

(a)



(b)

Plate 2.1 Examples of agricultural practices within the Yobe valley: (a) a Fulani herdsman rearing cattle; and (b) a tubewell-irrigated vegetable plot.

2.4 Geology.

Northern Nigeria comprises mainly crystalline Basement with Cretaceous continental and marine sediments in elongate basins, Tertiary continental sediments and Quaternary deposits. Across all components are alluvial lowlands associated with major rivers.

Figure 2.1 also includes a simplified surface geologic map of the Yobe basin and its surroundings. The materials within it may be grouped into three principal hydrogeologic categories: 1) Crystalline Basement Complex of Pre-Cambrian age; 2) sedimentary Chad Formation of Tertiary and Quaternary age; and 3) alluvium and aeolian sands of Quaternary age.

The Basement Complex, including igneous and metamorphic, underlies the Chad Formation. It is exposed in the SW edges of the basin and is encountered at increasing depths eastward towards Lake Chad. The following descriptions of the Basement rocks are given by Hazell et al (1988). The crystalline rocks of Northern Nigeria are mainly Pan-African migmatites, gneisses and granites, a few charnockites and amphibolites; some Jurassic granites and volcanics; and occasional younger basalt dykes and flows. They have been, to a greater or lesser extent, affected by movements, the most important of which is the Cretaceous plate separation, which followed-and was followed by-repeated movements along the same or parallel axes. The consequent development of steeply dipping joints has been followed by weathering; the depth of weathering naturally varies with the age and amount of movement, but often reaches up to 50m in depth.

Remnants of lateritic surfaces, with underlying kaolinite, cover the weathered zone in some places. In addition there is a variable layer of superficial aeolian drift up to several metres thick. In the zone of weathering, below aeolian and laterite surfaces, recognizable layers include those where

a) decomposition is complete,

b) only quartzite crystals are recognizable,

c) crystal forms are intact but feldspars are kaolinized,

d) unaltered rock has slightly dislocated crystals,

e) fresh rock.

The Plio-Pleistocene Chad Formation extends beneath the basin plain to form the base of the alluvium succession, and uncomformably overlies the Basement. It generally thickens downdip from the outcrop area, and the dip of the unit becomes greater towards the axis of Lake Chad. It attains its maximum thickness near Lake Chad and thins irregularly but gently towards its margin. The sediments vary not only in thickness and age, but also in texture. The Chad Formation is heterogeneous in its sediments as a result of the spatial and temporal variations in depositional environment. The result is a complex sequence of lensoidal strata with varying extensions and thickness, consisting mainly of clays and sandy clays of a range of colours interbedded with coarse deposits (sands and gravels) occurring at various levels (Barber, 1965). Lacustrine clays are predominant towards the centre of the Chad Basin and fluviatile sands, grits, and gravel occur near the margins. The complex structural aspects combined with vertical and lateral facies variations make prediction of lithology and other properties extremely difficult.

The upper unit in the basin is the recent (Holocene(?)) alluvium. The alluvium overlies the Chad Formation and consists of fluviatile deposits which are associated with the modern drainage lines of the Yobe River and its tributaries. The Yobe River and its tributaries are classical underfit rivers (past discharges exceeded present discharges), and occupy only a small portion of the valley, meandering across the top stratum of the alluvium.

The Yobe valley alluvium is poorly sorted, and the sediments comprising it include unconsolidated gravel, coarse and fine sands, silts and clays. Bedding of the sequence is also complex; lenses of sand of varying thickness and lateral extent alternate with less pervious materials or clay, or may be abruptly truncated by subsequent layers of sand and clay (du Preez and Barber, 1965; Hazell et al, 1988). Although cross sections of the alluvium show local variability in the sediments' character, the general relationship is of coarse sand and gravel in the substratum overlain by a more variable topstratum of finer grained sediments.

The nature of the alluvial deposits reflects the nature of the older deposits from which they were derived. Inspection of the surface geology in figure 2.1 shows the range of possible erodible deposits which occur in the Yobe basin and its surroundings. The valley sediments might have been formed as a result of the erosion and deposition by the headwaters of the accumulated surficial deposits of aeolian silts and clays, and the deeply weathered crystalline soils on the Jos and Bauchi plateaux. The coarse facies of the alluvium (coarse sand and gravel) were derived from the weathering of the coarse-grained Older Granites of the Jos and Bauchi Plateaux. Silty and clayey facies of the alluvium were derived from the weathered feldspars of the crystalline Basement and aeolian silts among others.

Hazell et al (1988) combined drilling and geophysical data and delineated the strata of the Jama'are River valley alluvium. They found the sediments comprising the alluvium to be variable and the Jama'are valley contained buried channels. The presence of the buried channels suggests that the sediments were laid down in much wetter conditions prior to the amelioration of the climate leading to the present. A probable sequence of climatic changes has been suggested by Pullan (1962) for the area west and south of Lake Chad. The sequence includes a drier period followed by a wetter period than at present, when the rivers were much larger and occupied wide channels.

The coarse facies of the alluvium (sands and gravels) were transported, deposited and reworked during the period of high river discharges accompanying wetter climatic conditions. Large river discharges also cut deep channels in the older deposits which were subsequently infilled with fluvial deposits during periods of low river flow. As the climate became drier and the water velocities in the rivers decreased, the coarse particles were replaced by finer materials such as silts and clays. The continuous process of cutting and infilling produced both the variations in the sediments' character and the complex buried topography.

Changes in strata and their erodibility also control both the location and width of the floodplain. One of the most significant changes occurs east of a line running roughly from Kari to Kano (figure 2.1) where the hard Basement rocks upstream are replaced by a relatively soft Chad Formation downstream. The line between Kari and Kano roughly represents the SW limit of the Chad basin sediments. This resulted in greater freedom of movement to the rivers downstream. As a result the width of the floodplain deposits widens rapidly from 1-2 kilometres upstream to up to 14 kilometres downstream. The part of the catchment underlain by Chad Formation sediments is small, and rainfall is low, compared to the upper part of the catchment underlain by the Basement Complex rocks, where rainfall is greater. Hence the depositional environments in the Chad basin differ from those in the upper more enclosed parts of the Yobe basin; the alluvium diminishes in grain size northeastwards.

Alluvium on the crystalline outcrop has pebble size particles but northeastwards gravels seem to disappear and sand size diminishes (Water Surveys, 1986). In addition, reduction of gradient and general fining of sediments downstream have led to significant increase in channel sinuosity (medium to high sinuosity). The most likely source of additional silts and fine sands downstream is from material transported by wind erosion north of Yobe floodplain (aeolian sands from the Manga grassland and elsewhere(?)). Such material could easily be tranported by wind into the Yobe floodplain and deposited in a predominantly fluvial environment.

The aeolian sands in the Yobe valley occur as a series of parallel dunes trending WSW-ENE. The ancient dune deposits occur over most of the area except where they are eroded by current drainage, while the modern dunes are mainly developed to the north of the Yobe River (for example the Manga Grassland dune fields).

Both the ancient and the modern dunes are unconsolidated. These strata, which are buff to orange, fine to medium grained micaceous quartz sands are typically 4-8m in thickness and contain iron-oxide nodules in places (Shultz, 1976). Little is known about their water resources.

2.5 Hydrogeology

Groundwater occurs in the Basement Complex either in the weathered mantle or in the joint and fracture systems in the unweathered rocks. The presence of groundwater in any given area therefore depends on whether a sufficient thickness and extent of decomposed material is present to provide a reservoir or, whether joints and fractures are present in the fresh rock. Storage is often the ultimate limitation on the yield of a borehole. The decomposed mantle is often too thin to harbour large quantities of water and is usually too clayey to be highly permeable. The joints are normally too poorly developed to compensate for the inadequacy of the zone of weathering overlying them (du Preez and Barber,1965). However, reservoirs of estimated dimensions of 50m depth, 40m surface width, and less than 200m length have been exploited by mechanical-pumped borehole (Hazell et al,1988).

In most cases the aquifer is unstable, clay-free only immediately above the bedrock, and the permeability of this material is quite variable and largely unpredictable (Beeson and Jones,1988). Little is also known about the porosities of weathered crystalline rocks. Under the Yobe plain the Basement Complex forms the effective lower limit of the groundwater reservoir.

The water bearing properties of the Chad Formation have been studied near Lake Chad (Barber and Jones,1960; Barber,1965; du Preez and Barber,1965). The following descriptions of the groundwater system of the Chad Formation are taken from Barber (1965) and du Preez and Barber (1965). Groundwater occurs in the sands of the Chad Formation under water table conditions, in perched aquifers, as confined water and as semi-confined water. Water occurs under water table or semi-confined conditions in the whole of the saturated section of the Chad Formation in the west and south and in the saturated portion of the top 60m to 120m of the formation in the northeastern part of the basin. The sediments occupied by the groundwater consist of intercalated lenses and beds of gravel, sand, silt and clay. As clean sands and gravels are common in the sequence of the western part of the basin, high yielding wells and boreholes are usually encountered. In the east, the sand lenses are less common and some boreholes have been drilled without encountering a sand lens or bed in the zone of normal groundwater.

Three zones of the Chad Formation sands containing groundwater have been clearly recognised below Maiduguri town and have been named the Upper, Middle and Lower zone Aquifers (Barber and Jones,1960).

The upper aquifer is the common source of water tapped by hand dug wells and boreholes in the basin. This aquifer occurs at about 30m to 100m below the land surface and is overlain by silty clay. The aquifers consist of lenses of fine to coarse sands intercalated with clays and silts. Individual lenses are not thick, rarely exceeding 5m. This zone apparently does not extend far to the north, east or west of Maiduguri but it probably persists for some kilometres to the south and southeast. During the 1950s and 1960s water was being pumped from this aquifer for the Maiduguri town water supply at a rate of about 1640 m^3/d and has certainly increased steadily up to the present day.

The middle aquifer has a large lateral extent and provides most of the borehole supplies. It is mostly confined or semi-confined by a continuous clay cover, and artesian flow occurs locally in some places. The top of this aquifer is situated at less than 100m below ground level in the western part, but dips sharply to approximately 200m near Geidam and from there dips gradually to 325m near Lake Chad (IWACO, 1985). This aquifer consists of fine to coarse sand and occur as interconnected lenses, intercalated with clays and sandy clays. Little is known about its thickness because most of the boreholes penetrate it only partially. There is yet no direct evidence of the safe yield of the system, but it is thought that the amount of water now being taken is released as a result of compression of the aquifers and confining strata in response to pressure relief.

Aquifers encountered in this zone have transmissivity of about 200 m^2/d and may have materials of uniform grading (Shultz,1976). Figure 2.3b shows grain size distribution curves for borehole samples obtained for this unit at several locations upstream of the present study site.

The lower aquifer is not considered here.

Though the alluvium is largely undeveloped, studies in the upper parts of the basin (Diyam, 1987; Water Surveys, 1986), show that the alluvial deposits are predominantly arenaceous, and where they are saturated, they make good aquifers.

The most abundant material is sand. Sands deposited by the rivers are fine to coarse grained, poorly sorted, commonly cross-bedded, and consists of angular grains of quartz; though lateritic and feldspathic materials are common. Figure 2.3a contains grain size distribution curves for samples obtained from boreholes located along the Hadejia River valley upstream of the present study site.

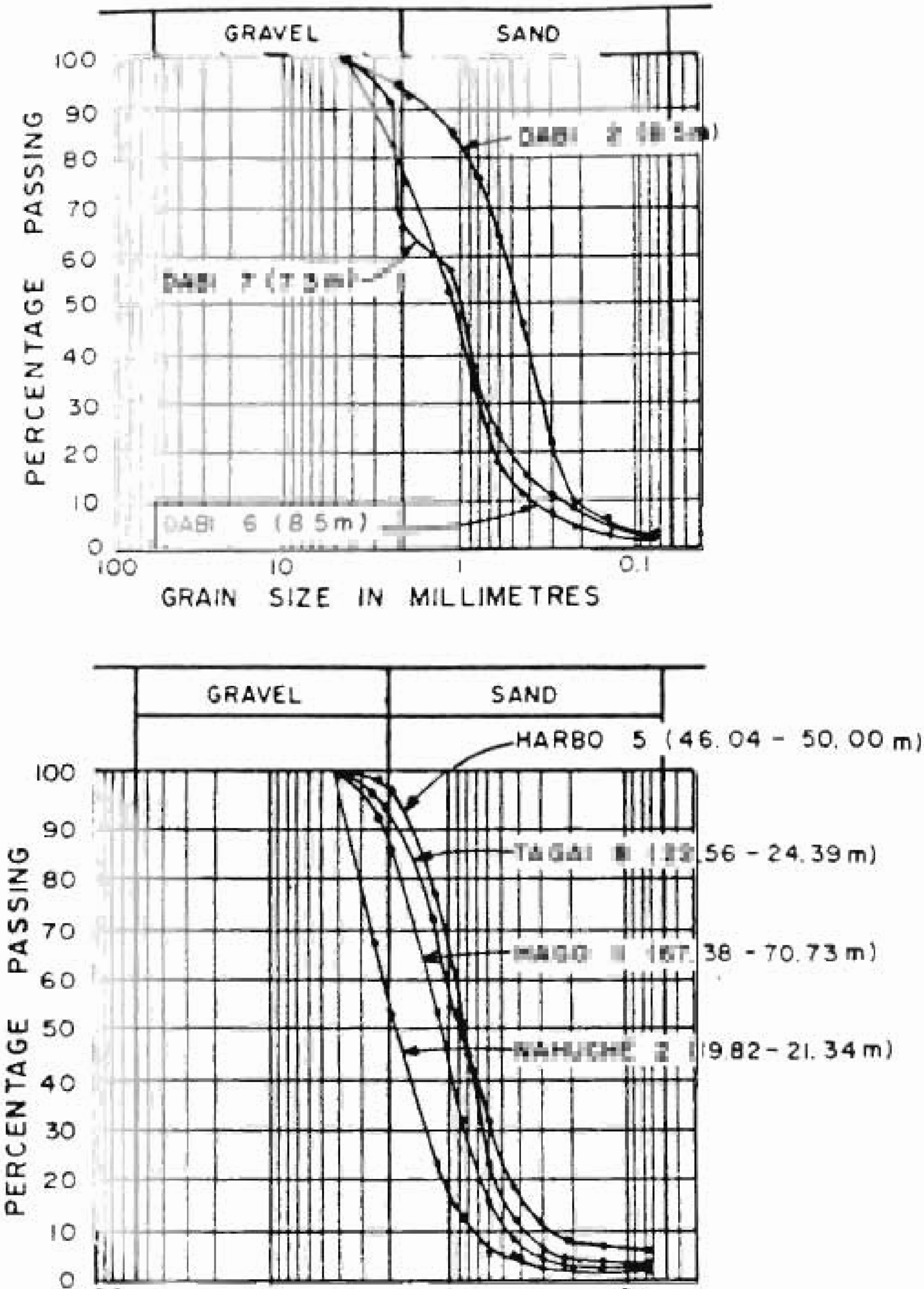


Figure 2.3 Grain size distribution curves for the basin aquifers (after Schultz, 1976) (a) alluvial aquifers, (b) middle Chad formation aquifers. Figures in parentheses are sample depth (m)

Gravels though at most places less abundant than sand, are predominantly of quartzite materials. They make the best aquifers, because it is easier to construct and develop wells perforated in gravel than in both sand and gravel.

The thickness of the aquifer is not well defined and varies along the basin. On the crystalline outcrop the thickness is in the order of 10-30m with an average value of 15m, but the maximum thickness encountered near Gashua Town is less than 20m.

There are few measurements of permeability of the alluvial aquifers. However, estimates made from grain-size data and pumping tests in Kano/Jigawa States (ie upstream of the present study site) by Diyam (1987), indicate a typical mean value of about 100m/day. The corresponding average transmissivity is $612m^2$/day, and the transmissivity ranges between 116 and $2000m^2$/day. The storage coefficient is between $2.0.10^{-6}$ and 0.17.

Contribution to groundwater in the alluvium is generally considered due to 1) lateral seepage from sustained high river stage; 2) overbank flooding and infiltration; and 3) less importantly from direct rainfall infiltration. The main natural discharge is effluent seepage to the river during low stage. Other discharges have been attributed to evapotranspiration, downward and outward flow to the Chad Formation aquifers (Schultz, 1976; IWACO, 1985).

2.6 Hydrology

Several rivers originate from the Basement Complex in the SW part of the basin and join together to form three major rivers namely: the Hadejia, Jama'are and Gana Rivers. Rivers Hadejia and Jama'are combine upstream of Gashua to form the Yobe River. Further the Yobe River joins the Gana River at Damasak and discharges into Lake Chad (Figure 2.1).

The tributaries of the Yobe River have a combined drainage area of about $60000km^2$ to Gashua and, are gaining streams at the upstream where substantial runoff occur as a result of high rainfall and impermeable strata. About 70% of the total runoff from the Basement area occurs in two months, August and September. Because of the sporadic nature of the rainstorms and the impermeable nature of the upstream strata, several flood peaks are recorded with the highest peak occuring in late August.

In the lower part the flood flow is highly attenuated because of the storage effect of topographic units such as backswamps and inter dune depressions. The flood peak in the lowland catchment is not a function of daily or storm peaks from the uplands, but rather a function of the total monthly or even seasonal inflow. Consequently a single flood peak moves steadily downstream across the lower part of the basin. The flood peak normally reaches Hadejia in September, Gashua by mid-October, and Lake Chad by January. After the flood peak passes, the flow recedes rapidly with virtually zero flow in the lower reaches, leaving behind swamps and pools of standing water.

The mean annual inflow to the lowland area upstream of Gashua (ie runoff from the Basement Complex) before the construction of Tiga dam for a ten year period of record (1964-1973) was 5410 million m^3 with a maximum of 8390 million m^3 in 1964 and a minimum of 2750 million m^3 in 1973, while the mean annual river flow at Gashua for the same period of record was 1380 million m^3 with a maximum of 2450 million m^3 in 1964 and a minimum of 630 million m^3 in 1973 (Sellars, 1981). Apart from the seasonal nature of the flow regime, the predominant characteristics of the Yobe River like most other rivers in a semi-arid environment (Balek, 1974; Hutchison and Midley, 1973), is the substantial flow losses in its lower catchment. The flow losses in the wetland areas between the Basement edge and Gashua amount to 70% of the total annual runoff and at Lake Chad the losses are up to 90% (Sellars, 1981).

CHAPTER THREE

THE YOBE BASIN: previous studies

3.1. Introduction

The first study of the water resources of the region comprising the present Borno and Yobe states was carried out by the Geological Survey of Nigeria (GSN) in the 1930's (Raeburn and Jones,1934), who reported on the geology and water resources of the Chad Basin. Other similar studies (Barber,1965; du Preez and Barber,1965) were also carried out more than 30 years later by GSN which also reported on the geology and water resources of the Chad Basin. In all of these studies, however, little attention was given specifically to the Yobe River basin itself.

In the last 18 years some studies have been carried out specifically to describe the land and water resources of the Yobe River basin. These studies include Shultz (1976), IWACO (1985) and Water Surveys (1986). A brief review is given in this chapter on the findings of some of these studies.

3.2 Shultz (1976)

Shultz International Limited undertook a study of the Hadejia-Jama'are River basin over a period of one year (1975 to 1976). The study was initiated by a technical agreement signed between the Federal Government of Nigeria and the Government of Canada in March 1973. The overall object of the study was to examine the Hadejia-Jama'are River basin's surface and groundwater resources, the potential for developing water storage and regulation facilities, and the potential of the land resources for irrigated agriculture.

Shultz's findings are reported in six volumes entitled "The Hadejia River Basin Study". The volumes are subtitled: A) Development Potential; B) The Environment; C) Soil Survey; D) Agriculture; E) Water Resources; and F) Legal Rights.

Because the present study aims to address some aspects of the basin's water resources, only the water resources volume is reviewed here. Even in this volume though, only those sections which are directly related to the present study are discussed. For example, sections highlighting criteria for selecting suitable dam sites and irrigable lands are not included.

The water resources study covered the catchments of both the Hadejia and Jama'are Rivers and the Yobe River as far as Gashua. Studies undertaken included the examination of the basin's stratigraphy; the development of a mathematical surface water model of the basin; and determination of the aquifer hydraulic properties and regional groundwater outflow.

The basin stratigraphy was examined both on a local and "regional" scale, for the Chad Formation sediments and on a local scale for the river alluvium, using lithologic logs obtained from exploratory drill holes and existing water supply boreholes. Locally the

1) lithology of the river alluvium was examined from auger holes drilled at three selected sites across the floodplains of the Hadejia and Jama'are River valleys. The Jama'are valley had two sections each with five auger holes, while a section with eight holes was established across the Hadejia River valley. At all the sections considered the river alluvium was found to contain 5 to 9m of medium to coarse grained sands overlain by a minimum of 3m of fine sands, silts, sandy clay and clays;

2) lithology of the Chad Formation sediment was likewise explored at one and two sections, each with three boreholes, across the Jama'are and Hadejia River valleys respectively. At these locations alternate layers of sands and clays of variable thickness, texture and colour were encountered; and

3) hydraulic properties of the Chad Formation aquifers were determined from pumping tests conducted at one and two locations respectively in the Jama'are and

Hadejia valleys. The aquifers exhibited confined characteristics during pumping and have moderate transmissivity (100 to 300m2/d). No pumping test was performed for the river alluvium.

The regional picture of the basin stratigraphy was viewed from a panel diagram constructed (using all the available logs) for a section of the basin from the upstream down to Gashua. This diagram shows the sedimentary sequence of the basin to be extremely variable in both thickness and extension and is difficult to correlate individual units basin wide.

A surface water model was developed and used to estimate elements of the hydrological cycle responsible for surface water storage changes in the middle section of the basin. The development of the model was based on the assumption that the flooded area in this section of the basin is a function of the system storage and the system storage comprised both surface and groundwater storage. With this assumption, a water balance analysis was carried out using the continuity equation and the system storage was obtained as cumulative inflows to the section minus outflows. The inflows to the system storage include rainfall and upstream river discharges, while river discharge at Gashua, evaporation from the wetted floodplain surface and open water area, evapotranspiration from recession crops, and regional groundwater flow comprise the outflows.

In performing the water balance analysis

1) the flooded section of the basin was conceptualised as an isosceles triangle with the river banks inclined to the horizontal and a floodplain storage model was developed to simulate changes in the flooded area with time which result from changes in the total amount of water entering or leaving the system;

2) the weighted monthly pan evaporation data were converted to open water and swamp evaporation using a pan factor of 0.7. The evapotranspiration from the

recession crops was determined using the Blaney-Criddle method and applied for the growing season;

3) the continuity equation was solved by setting a constant value for the regional groundwater outflow and adjusting the coefficients in the storage equation until the total storage change over the period of record (1964 to 1974) was equal to zero. That is the area flooded at the end of the study period was assumed equal to that at the beginning and subsequently used in estimating the initial system storage; and

4) the model was calibrated by comparing model estimates of areas flooded with those determined from aerial photography.

The average annual inflow into the Hadejia-Jama'are River system over 11 years (1964 to 1975) of record is 5350 million m^3. The average outflow past Gashua over the same period is 1350 million m^3. The annual open-water and swamp evaporation losses estimated from the water balance analysis varied from 930 million m^3 in 1973 to 3,760 million m^3 in 1964. The mean annual evaporation loss was 2,560 million m^3, about 65% of the total losses. The remaining 35% was assumed lost annually by regional groundwater flow away from the floodplain and by evapotranspiration losses from the shallow groundwater table in the vicinity of the flooodplain. The estimated annual groundwater outflow was 1440 million m^3, which is just over 25% of the total inflow (river inflow and rainfall).

Meanwhile, the groundwater outflow was also estimated using hydraulic gradients estimated from the potentiometric contour map constructed with water levels measured in a number of wells and boreholes in the basin, and the estimated permeability and aquifer thickness based on borehole logs. The regional groundwater outflow was calculated to be about 500 million m^3 per annum which is only about 30% of that obtained from the water balance analysis. Shultz (1976) attributed the apparent disagreement between the two estimates to additional storage depletion by phreatophyte evapotranspiration.

The water balance model approach, however, has the following short comings:

1) the groundwater flow pattern in the basin aquifer was based upon the hydrogeological concept assuming the aquifer to be one continuous, interconnected, homogeneous, and structurally undisturbed entity. The basin comprises sediments of variable composition and texture, and the hydrogeological implications of the presence of layers of low permeabilty (both on the surface and below) were ignored. The obvious bearing of such depositional elements on infiltration and groundwater movement in the basin has, thus, not been dealt with in the modelling; and

2) in order to fit the water balance model to the measured 1969 and 1974-5 flood extent data a constant groundwater recharge was chosen which would leave groundwater storage at the end of the 1975 equal to that at the beginnng (1964). Given the fluctuation in rainfall over the period (1964 to 1975), there is uncertainty in running a model with a constant recharge rate of 120 million m^3 per month regardless of the volume of river flow and area of flooding.

3.3 IWACO (1985)

In 1984 IWACO was commissioned by the Nigeria-Niger Joint Commission for Cooperation (NNJCC) to undertake a study of the Komadougou-Yobe basin water resources. The Komadougou-Yobe basin is the lower section of the Hadejia-Jama'are-Yobe River basin immediately downstream of Gashua to Lake Chad. IWACO, for the purposes of their study, termed this part of the basin the Master Plan Area (MPA).

The study reports, entitled " Study of the Water Resources in the Komadougou-Yobe Basin", are contained in seven volumes; a general report and six technical reports. The technical volumes are subtitled a) Basin Definition and Geomorphology; b) Climatic Analysis; c) Hydrology and Soil Salinity; d) Water Use; e) Groundwater Resources; and f) Surface water resources.

For the reason given earlier in section 3.1, only the water resources (surface and groundwater) volumes are reviewed.

The water resources study comprised:

a) a general review of the geology and hydrology of the Hadejia-Jama'are-Yobe basin;

b) an assessment of the processes that influence the basin's hydrological regime;

c) the development of a surface water model for the section of the Hadejia-Jama'are-Yobe basin underlain by the Chad sediments; and

d) estimation of groundwater recharge.

The hydrological data analyses included the estimation of the annual rainfall volume and runoff coefficients for the individual upstream sub-catchments. Also both the temporal and spatial distribution, and the occurence frequencies of the river discharges were examined. In addition upstream river discharges were related to downstream river discharges through regression analyses.

IWACO's model is a water balance model and in some aspects similar to that developed by Shultz. The two models are similar in the following respects:

1) the storage-flooded area submodel used by IWACO was adopted from Shultz;

2) a procedure similar to Shultz's was used to estimate the amount of rainfall falling on the flooded area and the evaporation from it; and

3) similar water management scenarios were evaluated for the middle section of the basin.

IWACO's modelling approach differed from Shultz's in the following respects:

1) IWACO's model comprises both the middle and lower sections of the Yobe basin;

2) the basin was schematised into 25 subsections and a water balance analysis was performed for each unit by taking inflow to a unit as the measured discharge from stations located at the upstream or outflow from a station(s) immediately above the subsection under consideration ;

3) evapotranspiration by phreatopytes (but not by recession crops) was incorporated into the model by performing a water balance analysis of the root zone with inflows from percolation from the wetted area. A maximum storage of 1000mm was maintained in the root zone and percolation in excess of this was assumed to go into groundwater storage;

4) the model was calibrated using average river discharges. Model parameters were adjusted until calculated discharge for a given station fitted the measured value;

5) the model was developed mainly to assess the effects of changes, both natural (climatic variations) and man-made (such as dams), on the hydrological regime of the basin; and

6) groundwater recharge estimated from the modelling study was considered too rough to be valuable.

IWACO (1985) estimated recharge rate for the MPA by assuming a uniform aquifer transmissivity ($100m^2/d$) for the aquifer they termed the "upper aquifer", and used the piezometric map shown in figure 3.1. The total recharge, which includes river recharge and lateral flow from adjacent aquifers to the "upper aquifer", was estimated to be about 21 million m3/year. This recharge estimate when re-calculated as a depth over the MPA ($23600km^2$), the value obtained is 0.9mm/a.

The estimate was made on the assumption that rainfall does not contribute to recharge.

However, in making the recharge estimate IWACO was not explicit on the following:

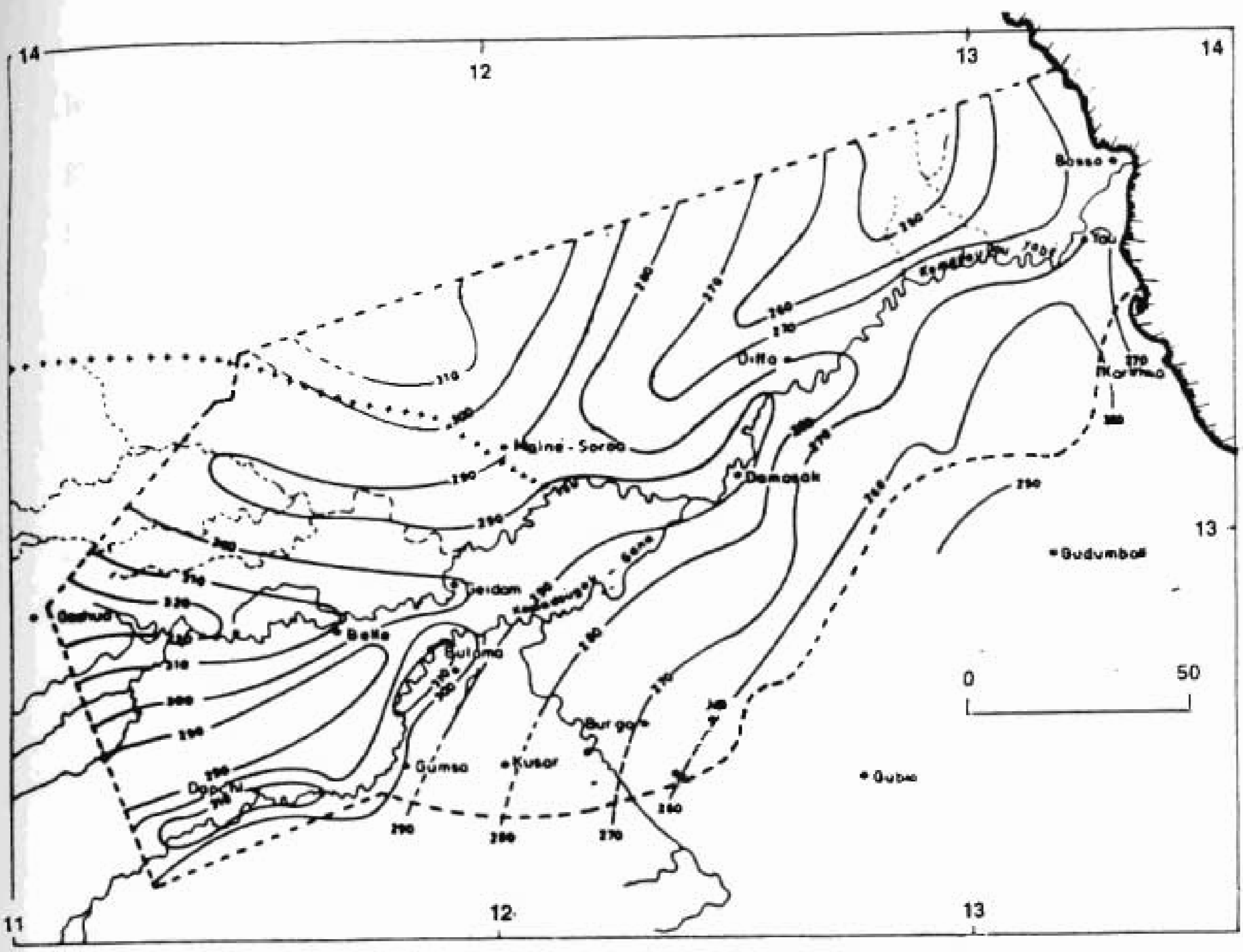


Figure 3.1. Piezometric contours of the upper aquifer (After IWACO, 1985).

1) Classification of the basin aquifer systems: IWACO divided the basin sediments into three individually extensive zones and named them the upper, middle and lower Chad Formation aquifers. The upper aquifer was considered to comprise up to 50m or more of the basin sediments.

Recent alluvial deposits of up to 30m maximum thickness occur along the modern drainage lines of the river systems (Shultz,1976; Hazell et al,1988) but no clear distinction was drawn between this unit and the older Chad Formation deposits. Thus IWACO's upper aquifer includes both the alluvial aquifer and other aquifers at depths greater than 30m. This raises the question of which aquifer the water levels used in constructing the contours shown in figure 3.1 were obtained.

Invariably uncertainty in aquifer stratigraphy and well perforation depth precludes grouping wells according to the aquifer being tapped, Yobe alluvium or Chad Formation. It is probable that the steepening of the contours at some locations (seen in figure 3.1) could be due to difference in water levels corresponding to aquifers at different levels and those caused by hydraulic barriers such as intercalated clays which commonly occur in the basin sediments. In the absence of detailed information on facies changes, aquifer thickness variation and resulting changes in hydraulic conductivities, the observed piezometric surface could, in the best case, outline only general trend; and

2) data points and ground level data: the points at which the water level measurements were made, and time of the year the data were obtained were not given; sampling distribution in space and time affects contouring. It is also unclear whether detailed topographic survey was undertaken to determine the ground surface elevation at the measurement points. Inadequate and unreliable ground surface elevation data can significantly affect the contours produced.

3.4 Water Surveys (1986)

This study was prompted by the increased development of shallow alluvial groundwater of the major rivers (namely the Jama'are, Gongola and Dingaiya River systems) for small scale irrigation in Bauchi state. Alluvial groundwater exploitation in Bauchi State started in 1982 and was encouraged and monitored by the Bauchi State Agricultural Development Programme (BSADP) which operated teams to jet down washbores. These initial developments were, however, carried out randomly and little information existed then on the potentials and sustainability of the groundwater resources. In 1984 the World Bank, the financier of the BSADP, recommended that studies be carried out to quantify shallow groundwaters and investigate abstraction techniques to ensure the safe and useful development of the shallow aquifers.

Following this recommendation Water Surveys Group (Consultant Water Engineers) were commissioned in 1985 to carry out surveys to:

a) identify and delineate the most favourable areas for development of alluvial lowlands;

b) determine the aquifer characteristics and to find out how much groundwater is available for small scale irrigation; and

c) identify the most suitable irrigation techniques for the various types of aquifer.

Because the Jama'are River is a major tributary of the Yobe River, and the fact that its valley alluvium was studied in detail by Water Surveys, the discussions that follow relate to the Jama'are valley sediments.

Water Surveys (1986) partly delineated the alluvial aquifers by classifying the river systems in terms of sinuosity and sediment load. The present and fossil channels were classified according to their sinuosity pattern at various stages using aerial photographs complemented by Landsat false colour satellite image. Consequently a basic distinction was drawn between low sinuosity (L-S) and high sinuosity (H-S) or meandering channels.

Classification of the flood plain surface features was followed by the determination of the subsurface lithology using geophysical techniques and drilling. The technique used was to estimate the conductivity of type layers, find out the most common layer configurations and thicknesses and establish the expected responses from computer models and compare with field electromagnetic (EM) profiles. The theoretical EM profiles were, where appropriate, adjusted to give best fits to the field data. In addition, anticipated EM responses corresponding to Vertical Electrical Soundings (VES) interpretations at particular locations were also calculated and compared with field results. The type sediment sequences established were finally compared with field VES sections and drill logs.

The above approach was generally found by Water Surveys to be suitable and cost effective for exploring sand and gravel units covered by a thin conductive surface layer in both the present and past river channels. This aspect of the study has been published by Hazell et al (1988). They concluded that " a combination of resistivity soundings, especially to help calibrate the electrical section, and EM traversing at carefully selected separations aided by computer modelling gives best promise for groundwater exploration in the alluvial environment". The EM data interpretations, however, became extremely difficult in the presence of thick conductive surface unit such as clay pans, and the results obtained in this situation are thought to be less accurate.

Three main types of aquifer targets were defined; type A, the present and past channel deposits (low sinuosity channel), type B (the meander point bar sands), and the sub-pan or relic meander point bar sands (type C). Buried channels containing thick coarse deposits were identified in both the upper and lower reaches of the Jama'are River. The sediments comprising the aquifers are poorly sorted and they include fine, medium and coarse sands/gravels, 10 to 15m thick. Pans and interchannel flood basins are less common upstream (Bunga), but well developed downstream (Mellemdige), and the clastic materials associated with them are up to 7m thick. Refer to figure 2.1 for the locations of these towns.

Pumping tests were conducted at six locations and the results indicate that the Jama'are valley aquifer has a wide transmissivity range (300 to 2000m^2/d) and behaves as an unconfined aquifer during pumping.

The influence of river flow on aquifer water level was monitored using piezometers installed at nine locations, from Bunga northwards, within the present river channel. Water Surveys (1986) observed little time lag between river flow and aquifer response in the upper reaches (within the Basement Complex) , and aquifer water level increase was immediate. However, a lag time of a few days was observed at the fringes of the

Basement and subsequently increased to about 25 days in piezometers located over the Chad Formation sediments.

Based on the above observations Water Surveys concluded that "recharge of alluvial aquifer is dominated by river flow. There is good hydraulic continuity between river flow and aquifers and under present climatic conditions and river regimes, completely depleted aquifers (to the suction limit) will be completely recharged annually". However, it must be realised that this conclusion is valid only for parts of the aquifer beneath the river channel since piezometers installed in a river channel measure the head changes in the channel aquifer, but can not reveal how aquifers beneath the adjacent flood plain react to the river flow. As a result the river-aquifer interaction was only partially determined. A complete definition of river-aquifer interactions requires the determination of the aquifer water level conditions (confined or unconfined) beneath the flood plain and how river stage changes affect water levels in piezometers installed normal to, and at distances from the river.

3.5 Diyam (1987)

This study was carried out along the Hadejia River valley in Kano state. It was also brought about by the need to develop shallow groundwater for small scale irrigation. The study was initiated by the Kano State Agricultural and Rural Development Authority (KNARDA) which commissioned Diyam Consultants in 1986 to carry out studies on the Hadejia valley alluvial groundwater. The objectives of the study were similar to those in Bauchi state. These include

1) delineation and identification of areas of alluvial aquifers; and

2) determination of aquifer hydraulic properties.

Diyam (1987) made a general review of the geology and water resources (both surface and underground) of the Yobe basin and its surroundings. The review was followed by an inventory of wells and boreholes within the flood plains of the major rivers (namely Hadejia and its distributaries), and areas with dry season groundwater levels less than 6m below the surface were identified. Five areas were identified along the Hadejia River and its distributaries while a sixth area was located within the lower end of the Jama'are valley in Kano State.

Fifty five (55) resistivity soundings were carried out using the Schlumberger configuration to identify depth to water table and lithology. This technique was, however, found by Diyam not to give satisfactory results in the Hadejia valley alluvium. The interpreted VES profiles do not always agree with the available drill logs. As a result Diyam stated that "even with the benefit of some knowledge of the geology and groundwater level it was not possible to identify the water table or lithological interfaces at most of the test sites".

However it is worth noting that from the results included in Diyam (1987), it appears that the VES data were interpreted independently, and then compared with the lithologic logs. Unconstrained interpretation of geoelectric data often give ambiguous results because of electrical equivalence, and so geoelectric survey is neither unique nor diagnostic by itself. Clearly, there was the need to constrain the VES interpretations with the drill log data at least at some locations to establish the resistivities corresponding to different units of the geologic profile.

Aquifer hydraulic parameters were determined at eleven locations using pumping tests results. The tests show that the Hadejia valley aquifer has a wide transmissivity range (116 to 2122m^2/d) and behaves as a confined aquifer during the tests.

Water levels were monitored in 78 selected wells to determine aquifer response to changes in river stage and to estimate recharge. The water level data included in Diyam (1987) show the following:

1) localised water level increase was observed with only two or three wells responding to river stage changes in each of the defined areas;

2) water levels in all the wells monitored, except in the area near Nguru, were more than 3m below the surface and showed negligible increase during the wet season;

3) in the Nguru area, 3 out of the 14 wells monitored have water levels within 3m of the ground surface and the average water level increase in the wet season was about 2m;

4) in all the areas the observed water level increase was negated by a rapid decline following the wet season recession; and

5) though 6m was taken as the limiting depth in defining the areas, more than 30% of the wells monitored have, even in the wet season, water levels deeper than 6m (in some wells deeper than 10m) below the surface. This observation brings in the question earlier raised, whether the aquifer monitored was the alluvial aquifer or the Chad Formation aquifer.

Recharge was estimated by multiplying the difference between wet and dry season water levels by the aquifer storativity. The average recharge calculated ranges between 195mm and 90mm. This average value assumes, of course, that the wells locations selected represented average conditions within a given area.

However, the criteria used by Diyam in choosing the aquifer storativity are not clear, and this is a crucial parameter in the recharge estimation. While their pumping tests results consistently gave a confined aquifer storage, they decided to use an unconfined storativity (15% specific yield) in the recharge calculations. The 15% specific yield used is several orders of magnitude larger than the storativity obtained from the pump test results, and, hence, the recharge volume calculated was an overestimate for an aquifer under confined condition. It was appropriate if Diyam had used the aquifer storativity obtained from the pumping test results since such tests are normally carried out in a study for the purpose of resources evaluations.

3.6 Other reports

Other reports on the water resources of the basin exist (NEAZDP,1990; Wardrop,1993), but are not considered here because they contain little (new) information different from those reviewed earlier.

CHAPTER FOUR

MATERIALS AND METHODS

4.1. Introduction

The research method adopted for the present study was a combination of several techniques, which are conventionally employed in the exploration and development of groundwater, and definition of its interaction with other elements of the hydrologic cycle such a river. All elements necessary for achieving the objectives of the present study were defined and quantified either from field data, or from previous studies; the former was obtained from instruments/equipment installed/used as part of the present study, while the latter was the subject of a desk study.

The techniques used include:

1) remote sensing applications;

2) resistivity depth sounding;

3) drilling and monitoring piezometer network installation;

4) river stage measurements;

5) surveying;

6) pumping test; and

7) conceptual modelling.

4.2 Desk Study

Mapping and Airphotography

A remote sensing technique comprising aerial photograph interpretation supplemented with Landsat false colour composite imagery was used to define the different physiographic units within the Yobe basin and consequently delineate the lateral extent of the floodplain.

The use of remote sensing data for groundwater studies has been demonstrated by some of the early works with Landsat imagery. Short (1973) reported that in regions with sparse vegetation where distinct features such as outcrop patterns occur, geologic reconnaisance maps could be prepared from satellite imagery. Short stated that such maps are superior to ground based geologic maps because contacts could be delineated better in the overview. Zall and Russell (1979) combined satellite imagery and aerial photographs and successfully delineated permeable rock aquifers in east and west Africa. They found that using landsat imagery in the reconnaisance stages of their study substantially decreased the cost and increased the efficiency of the groundwater exploration operations. High potential groundwater regions delineated from the imagery were further substantiated with aerial photograph interpretation, geophysical and ground surveys. Zigich and Kolm (1982) and Dene et al (1984) have also shown the applicability of the Landsat imagery in the delineation of buried glacial aquifers. Dene et al (1984) concluded that "the combination of remote sensing, geophysical techniques, and test drilling has proven useful for delineating some buried valleys in northeastern Kansas. Analysis of tonal patterns on spring Landsat imagery provides a quick and inexpensive first step for selecting field study sites".

Rahn and Moore (1979) delineated shallow glacial aquifers in the USA (eastern south Dakota) using Landsat imagery and concluded that: "Landsat imagery is a useful tool for groundwater exploration in a terrain where detailed geologic maps are unavailable and can enable the hydrogeologist to map shallow aquifers which have surface expressions with almost the same degree of accuracy as a conventional large scale field reconnaisance geologic map. The imagery provides a regional reconnaisance tool of considerable potential value, and in some instances regional patterns of aquifer distribution can be predicted with better accuracy than previously possible".

The combined use of aerial photographs and enhanced satellite imagery with ground control was also found to be efficient and cost effective in delineating alluvial aquifers in northern Nigeria (Water Surveys,1986; Hazell et al,1988). Features such as buried channels, point bars and ox-bows, classified by the aerial photograph interpretations as possible aquifer sources were substantiated by geophysical survey and drilling.

Prints of black and white aerial photographs and false colour composite Landsat imagery subscenes covering the study site were obtained and interpreted by conventional stereoscopic analysis and computer assisted image interpretation. The North East Arid Zone Development Programme (NEAZDP), Garin Alkali, Nigeria, provided the photographs for the study area. The photographs are on a scale of 1:25,000 and were acquired in 1990. The Landsat imagery is the Thematic Mapper bands 2,3 and 4 obtained for the study area by the Silsoe College Remote Sensing Centre, UK. The imagery was captured on 26/11/86.

Figure 4.1 is a subscene of the landsat TM bands 2, 3 and 4 imagery used. The contact between the floodplain and the adjacent uplands is quite distinct on this imagery. The

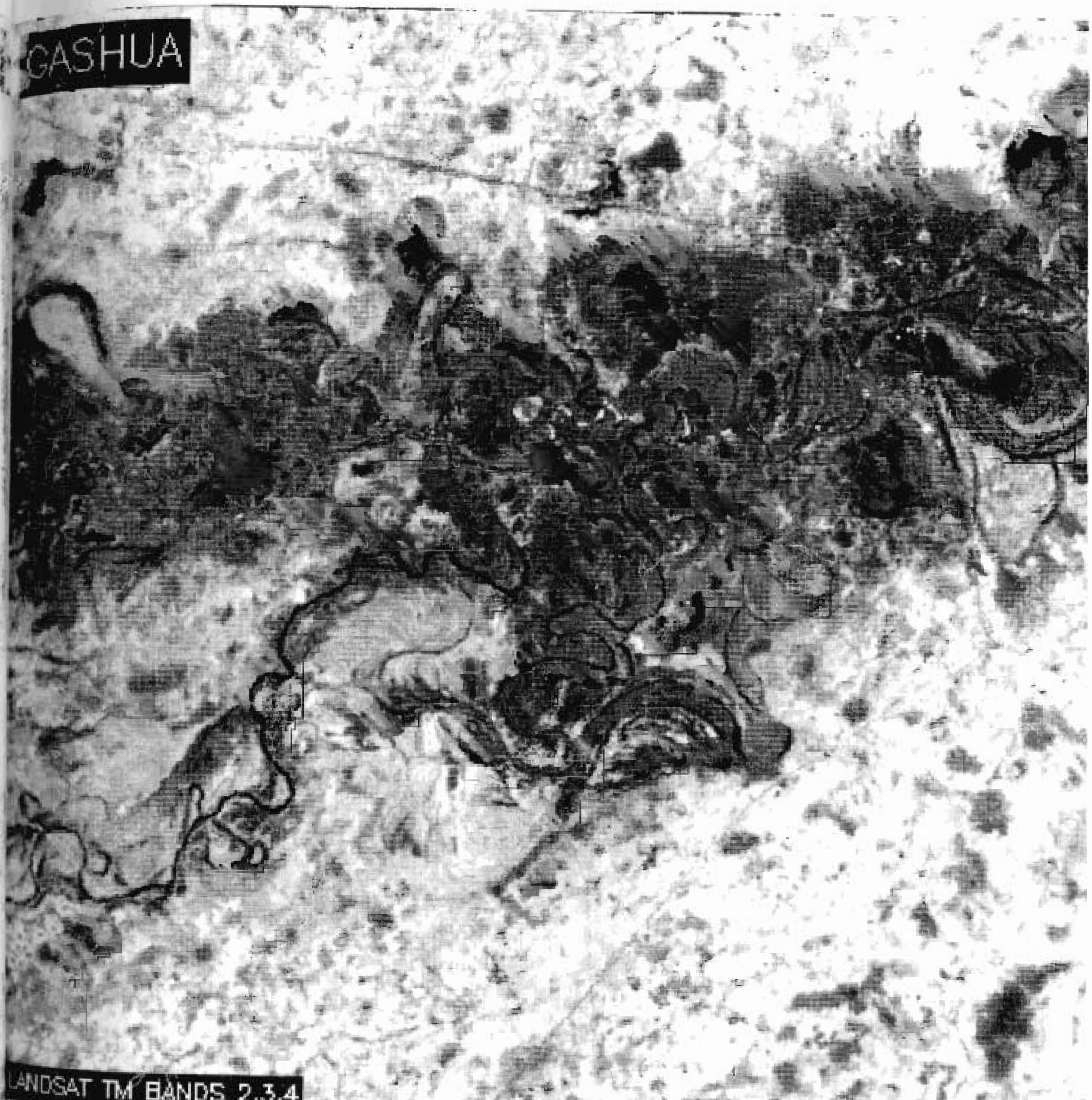


Figure 4.1. Computer enhanced colour composite image of (part) the Yobe Valley. The mottling patterns and river meanders are visible on this imagery. Image acquired on 28.11.1986.

dark area represents the floodplain and contrasts with the mottled aeolian covered surrounding uplands.

The floodplain has:

1) low infrared response. The floodplain has little vegetation and is mainly covered with clayey soils which are poorly drained compared to the aeolian soils of the upland;

2) elongated outline, caused by the orientation of the valley;

3) meandering river, oxbows and abandoned channels some of which are visible on the landsat imagery; and

4) serrated pattern along the edge of the valley, caused by river meanders.

Because the Landsat imagery was able to define the floodplain boundaries from the adjacent uplands at the early stages of the study, it eliminated the need to construct time consuming aerial photo mosaic for the initial "first look" phase normally required to get a feel of a project area. It also reduced the subsequent amount of photo interpretation necessary for the photogeological phase of the ground water exploration.

The mapping of the floodplain boundaries within the study site was carried out in the following ways:

1) Preliminary interpretations of photographs and the Landsat imagery were carried out in the laboratory and different physiographic units were identified. A subscene covering the study site was extracted photographically and the floodplain lateral extent (boundaries) was delineated, while exploiting the existing tonal and spectral characteristics exhibited by the surfaces of the floodplain and the adjacent uplands. The floodplain boundaries were traced directly on to a clear acetate overlay ;

2) correlation of the floodplain boundaries as defined by the imagery and the photographs with ground data. Detailed field surveys were carried out in respect of soils and land use patterns, and topography. Soil profiles were examined in addition to drill holes. Feasible boundaries for physiographic units, soil units, and land use were delineated. The field data were finally compared with the desk photo interpretation. In this particular case floodplain boundaries defined by photographs and Landsat imagery interpretations were well correlated with ground data; and

3) final preparation of the floodplain surface map with a cartographic elaboration. This included addition of legend and drainage patterns.

This three stage approach encompassing satellite data, aerial photographic data, and finally field checking has substantially aided this study.

4.3 Field Installations and Data Collection

Site Selection

The selection of the site at which the monitoring equipment were installed and field data collection carried out was determined by a number of factors. Some of these factors include:

1) Access: the presence of the Garin Alkali-Gashua road that traverses the floodplain has influenced the choice of the study site. Generally enormous practical difficulties exist in getting into the Yobe valley during the wet season after the floodplain surface has been inundated. The hard floodplain soils of the dry season is turned into mud and the entire floodplain is impassable to motor vehicles. Therefore there was the need to have a site which was relatively accessible;

2) Availability of facilities and materials: the site selected contained a gauging station at the Gashua Bridge which has been monitored for many years, and this facility was utilised for river stage measurements in the present study. In addition, the site was close to Gashua town where basic materials such as water needed for field installations were readily available;

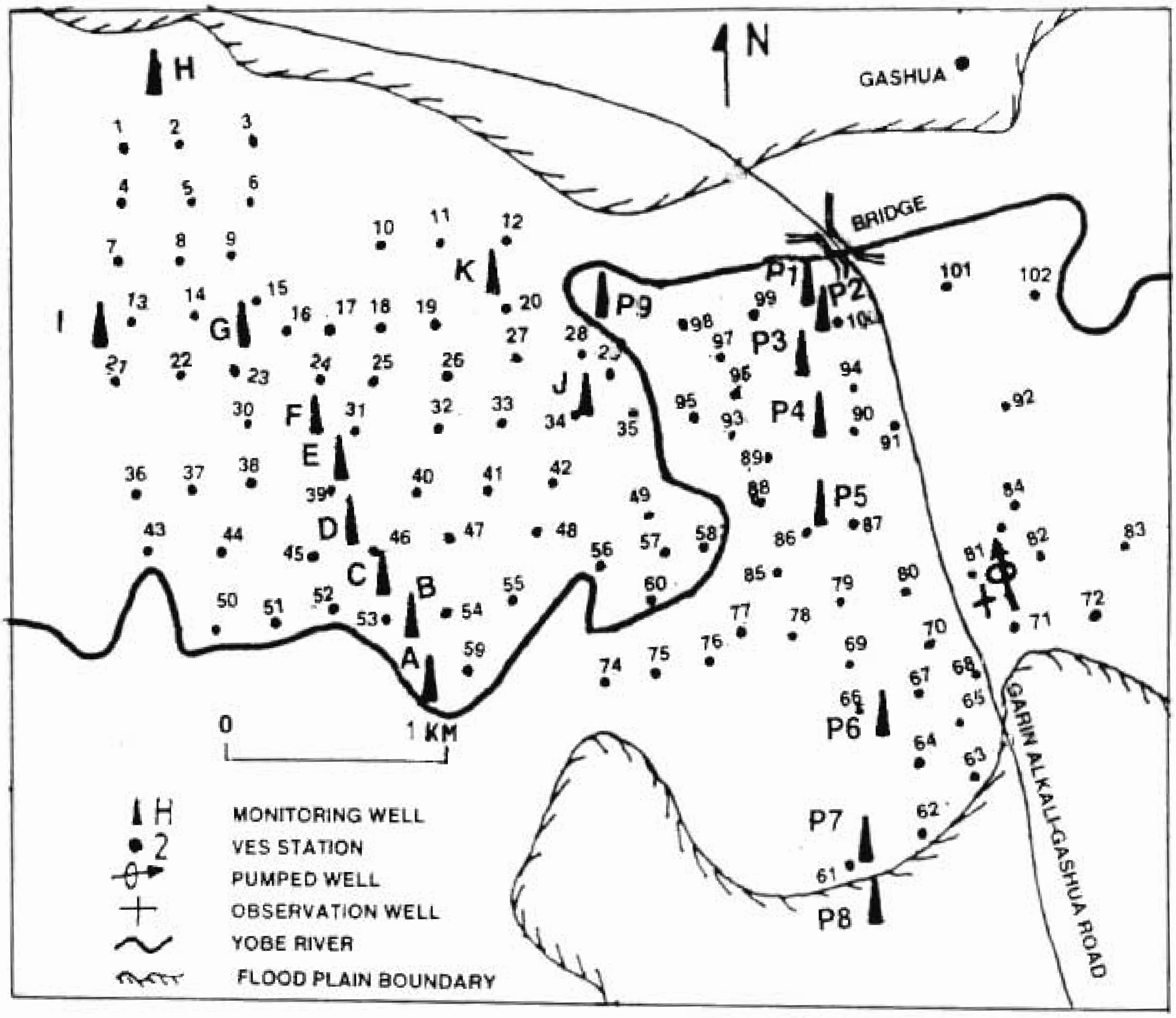


Figure 4.2. Map showing the location of geoelectric stations, monitoring piezometers and pumping well.

3) Security: measures to reduce damage and theft to the installed property was also a factor in deciding the site. Vandalism to the installed property was readily checked only because the site was close to the field station; and

4) Accomodation: there was the need to choose a site which was close to Gashua town where adequate accomodation was available.

The study site selected with field measurement points and installations is shown in figure 4.2.

Resistivity Depth Sounding

A geophysical technique was used to investigate the subsurface conditions within the Yobe valley floodplain.

Many geophysical techniques with different advantages are available (Driscoll, 1987; Fetter,1994); the one chosen for this study was resistivity depth sounding, which is also referred to as Vertical Electrical Sounding (VES). VES was employed because:

1) it is an inexpensive, fast and non-invasive technique that yields useful information about subsurface conditions;

2) it is well suited to the investigation of alluvial aquifers where various lithological units often have large resistivity contrast. In the Yobe valley floodplain, the primary targets of the VES survey are the top low resistivity clay layer and the underlying sand and gravel units documented by drill logs of wells located across the valley. The contrast in resistivity between the clay units and the underlying permeable sands and gravels made possible the determination of the alluvium thickness and extent; and

3) it has been successfully used by others (Water Survey, 1986; Hazell et al, 1988) in the region.

Resistivity sounding is a technique designed to examine horizontal or nearly horizontal geological layering and is widely employed in engineering investigations to measure depth to bedrock, and to determine the nature of the overburden and other ground conditions.

The use of direct current geoelectric soundings in ground water investigations is discussed in detail by Zohdy et al (1974). The method involves passing a direct current through the ground using a pair of electrodes and measuring the potential drop on the surface with another pair of electrodes. Electrode configurations are varied during the survey and apparent resistivities are measured at each sucessive interval until the target depth has been investigated. The apparent resistivities calculated at various electrode spacings result in a resistivity or vertical electrical sounding (VES) curve. The VES is interpreted to determine resisitivity values and respective thicknesses for individual layers at different depths using curve-matching methods (Zohdy et al, 1974) or computer programs, such as Offix (1988).

There are almost as many different opinions concerning the successes and/or failures of geoelectric resistivity methods as there are users. One of the reasons for unsuccessful geoelectric surveys is that apparent resistivity values may not unambiguously characterise major geological units (because of electrical equivalence). A study by Lennox and Carlson (1963) to determine the thickness of glacial drift over bedrock was unsuccessful because the two units possessed nearly identical electrical properties. Similarly, Dyck (1973) showed that the resistivity ranges of common glacial drift units such as sand, till, silt and clay tended to overlap one another. Extreme variability in the geology and a relatively deep water table were identified by Stollar and Roux (1975) as conditions that could also contribute to an unsuccessful survey.

Equally, however, there have been considerable successful geoelectric applications to maintain interest in the method. Page (1968) used the resistivity method to map buried stream channels in the Santa Clara Valley in the USA. Heigold et al (1979) outline a procedure for using electrical sounding to evaluate the water producing capabilities of granular aquifers. Huntley (1986) noting that electrical sounding measures a volume of sediment similar to that of an aquifer test, related aquifer permeability to measured resistivities. Barker (1988) attributes significant savings in manpower and increases in efficiency to resistivity sounding while using it for delineating subsurface features. Some other successful applications of electrical resistivity surveys include identification of a buried fault in the Bunker Hill groundwater basin in San

Bernardino, California (Park et al,1990), mapping the layered structure of a closed land field in Dupage County, Illinois (Carpenter et al,1990), and the delineation of coastal aquifers in Italy (Quarto and Schiavone,1994).

Three arrangements of electrodes are in common use (Fetter,1994). The dipole-dipole array has a pair of current electodes separated from a pair of potential electrodes. The same spacing is used between the current electrode as between the potential electrodes, and the distance between the electrode pairs is much greater than the electrode spacing. This configuration is particularly convenient for making electrical soundings which measure the changes in electrical properties with depth.

The Schlumberger array uses closely spaced potential electrodes and widely spaced current electrodes in a straight line. The potential electrodes are only expanded once or twice during a depth sounding to guarantee a sufficiently accurate voltage measurement. The main advantage of the Schlumberger method is that the data obtained are less affected by near surface lateral inhomogeneities (Flohlich,1973). One disadvantage of the Schlumberger technique is that the sounding curves are inherently discontinuous. This results from the need to periodically expand the separation between the inner potential electrodes to increase the signal level.

The Wenner array uses equally spaced electrodes. A current electrode is on each end. The main advantage of the Wenner method is the greater accuracy of the measured voltage and has characteristically high vertical resolution for horizontally layered media than the Schlumberger method (Barker,1979). The data obtained using this method, however, are affected more by near surface lateral inhomogeneities.

The Offset Wenner technique using multicore cables, has been devised by Barker(1981). This method is capable of reducing lateral variability effects and yet produces a Wenner apparent resistivity curve, but one of much better quality than a normal Schlumberger or Wenner sounding curve at the same site. The use of this method greatly reduces the effects on the sounding curve of near-surface lateral resistivity variations and allows checks to be made in the field on the reliability of the data obtained (Barker, 1981). In addition, the multicore cable method has the following advantages (Barker, 1988):

1) fieldwork efficiency is increased by using the multicore cable for simultaneous connection of all the current electrodes to the power transmitter. To change electrode position, a simple commutation unit (switchbox) is used. This leads to:

a) decrease in manpower-used with a modern digital resistance meter, the whole system is easily carried and operated by one man;

b) increase in productivity-a vertical sounding may be carried out in less than one hour and consequently, output may be raised to up to 10 or more soundings per day per man; and

2) spacing errors are eliminated-electrodes can only be connected to takeouts at correct positions on the cable.

In the present study the Offset Wenner Technique was employed by combining the BGS 256 multicore cable and switchbox with the ABEM SAS digital Terrameter.

The complete set of the BGS Offset Wenner system consists of two multicore cables (which are each 256m long) fitted with metallic takeouts at pre-determined distances, a number of electrodes, a switch box and cables to connect the system to a resistivity meter. The system uses a five electrode array with electrodes equi-spaced. However, only four electrodes are used at a time. Thus, for each electrode spacing a series of measurements can be made using different electrode combinations. The configuration of the Offset Wenner technique is shown in figure 4.3, while plate 4.1 shows the use of this technique in the field.

Two of these configuration D1 and D2 are conventional Wenner arrays but shifted by one spacing. Barker (1981) has shown that by taking the mean of these two readings a substantial reduction in the effects of lateral inhomogeneities can be achieved. The three remaining configurations A, B and C do not use the central electrodes but are a tripotential system. They allow interpolation between points produced by the Offset Wenner readings.

Three measures of error are possible from the readings. The first, the reading error, is a measure solely of the instrument, operator and electrode array. The sum of readings B and C must equal A, irrespective of the subsurface conditions. Thus any significant deviation from this equality indicates a malfunction in the system. Barker (1981) has given a single root mean square (rms) of 2% as the upper limit for the observation error. The second measure is that if subsurface conditions are laterally homogeneous, the two readings D1 and D2 will be equal. Hence the difference between the two readings is a measure of subsurface lateral inhomogeneity. The error measure, the potential ladder error, is generated when the sets of A, B, and C readings are used to interpolate between the offset data. This error is also a measure of subsurface lateral inhomogeneity. The upper limit for the offset error is 25% (rms).

A total of 102 VES stations were established across the study site (Figure 4.2); 60 were located north of the Yobe River, and the remainder south of the river. Most VES stations were located on an approximate grid of 256 by 256m which corresponds to the length of the BGS cables. The positions of the stations were determined using airphoto mosaics and compass, and marked with timber pegs. Maximum current electrode separation was either 64 or 128m; the majority of the soundings used a maximum of 128m. This current separation provided adequate electrode spacing for the target depth of 20-30m, the maximum anticipated depth to the clays underlying the sand/gravel aquifer.

The resistivity survey was carried out between November 1991 and February 1992 during which the floodplain surface was moist allowing easy penetration of the current electrodes into the ground.

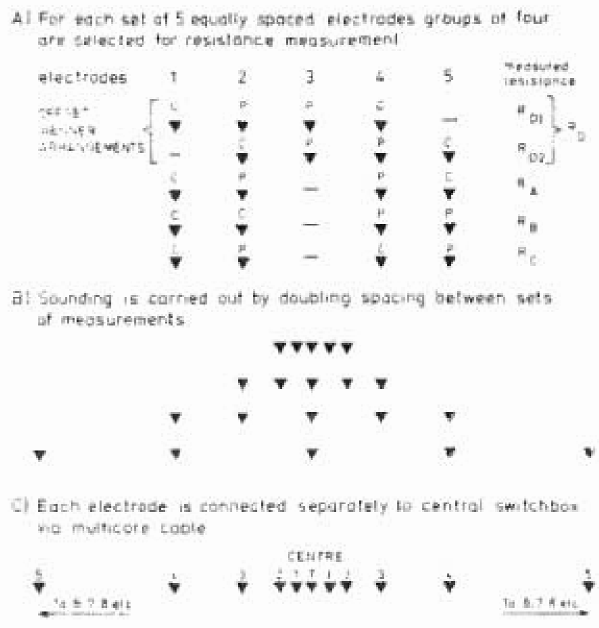


Figure 4.3 The offset Wenner sounding configuration (after Barker, 1981)

Plate 4.1 The use of the offset Wenner configuration in the field (a) the Barker cable with electrodes fixed into the ground; (b) the switch box (white); and (c) Terrameter (red).

A computer program, OFFIX (1988), which has both forward and inverse models was used to perform the quantitative analysis and interpretations of the VES field curves. The program automatically interprets the VES field data with regards to a user-supplied starting model. The starting model and the corresponding resistivity transforms are refined or modified by the program to obtain a best fit relation to the field data. As a final process, the program calculates layer thicknesses and resistivities, error closure statistics, and geoelectric parameters, and generates a geoelectric section for the measurement station.

Laterally homogeneous, isotropic and horizontal layers are assumed in the quantitative analysis. The fundamental problem generally encountered in the interpretational process is the derivation of an appropriate model consistent with available geological and geophysical information, within the limitations of the algorithm used. These limitations are related mainly to the number of layers and to the high CPU time required for computing a geoelectrical model. Therefore, an effective requirement was to start with a good initial model and adjust it by small changes in the electrical parameters.

A constrained modelling scheme was used at locations where geologic logs were available. This was done to avoid the uncertainties inherent in unconstrained modelling. Unconstrained modelling results in a non-unique solution and also resistivity and thickness of a single layer cannot be resolved accurately (Park et al, 1990).The geoelectric interpretation was performed at each borehole site by keeping known geologic thickness constant and varying the resistivities of the units until a good match was obtained between the model output and the field data. Resistivity ranges for the units were set as a guide and control for the inversions at adjacent stations where borehole logs were not available. The inversion results were accepted as a final interpretation when the model output was geologically and hydrologically reasonable; the output is similar to those at adjacent stations; and no wells were close enough to the sounding to correlate the logs with models. Otherwise, the inverse model was modified using forward modelling and constraints from logs.

Drilling and Drill log data

Geologic data at the study site were collected from borehole geologic logs and samples, and from a limited number of surface exposures, profile pits and trenches (plate 4.2).

Drill logs were obtained from Government agencies such as the Yobe State Water Board and the Chad Basin Development Authority. All geologic drill logs record qualitative grain size information, but lack consistent documentation of other supplemental data relevant to hydraulic properties including quantitative assessments of sedimentary structure and grain size.

The mud flush rotary method was used for drilling boreholes, 100mm in diameter, to a depth varying from 15m to 30m. This method was found to be very efficient for drilling in the Yobe alluvium. It was fast and any slumping in the hole was invariably cured by improving the viscosity of the drilling mud by adding more drilling substance. At least a hole was drilled every day. The greatest weakness of the method was the need to have a large volume of water for drilling. In this particular study a lot of time and resources were spent in transporting water to the drilling sites.

A portable trailer-mounted rotary rig, mechanically operated, and designed to be moved around using light weight vehicles, was used. A four wheel drive Land Rover vehicle was used in the present study. The body of the rig comprises a floor on which is mounted a petrol-engined power unit, a suction mud pump for circulating the fluid, a winch for raising or lowering the drill string and a mast from which the drill string is suspended. The drill string is made up of lengths of drill pipes (each 1.50m long) with the drill bit assembly (drill collar and bit) attached to the bottom. The top end of the drill string, the kelly, is a square section. All members of the drill string are connected

(a)

(b)

Plate 4.2. Clay thickness sampling

(a) Profile pit dug at a location in the study site; and (b) examining a section of the exposed river bank.

In both cases the clay thickness is greater than 1.0m.

by standard API taper thread joints. A swivel, which contains a bearing assembly, carries the complete weight of the drill string. This unit also has an entry for the drilling fluid passing up from the mud pump through the kelly hose and a suitable gland to control the passage of the fluid from the static swivel to the rotating kelly.

A polymer drilling mud (CMC (trade mark)) was used for drilling. The advantage of using this particular mud, like other polymers, was that it has a shorter life (about three days) than bentonite mud , after which it breaks down to a water-like consistency and is easily drawn out of the formation. In addition, it has only a weak thixotropic properties and releases solids much more readily.

The drilling fluid was mixed in a mud pit, which was prepared at each borehole location before drilling commenced. Immediately after mixing, the mud properties, density and viscosity, were measured using an API mud balance and marsh funnel respectively. The tests were also carried out routinely whilst drilling and circulating. 20 mud circulation pits were prepared. At each location the mud circulation layout has both settling and suction pits, which were 2.4x1.2x1.0m and 1.2x1.0x1.0m respectively. The pits were connected to each other by a circulation channel. The floors and walls of the pits and channel were sealed with screed.

Drilling at each location began with the installation of a length of conductor pipe to prevent erosion of the surface by the mud flow. The water swivel was then hoisted up the mast and the drill bit assembly screwed on to the kelly. The bit was lowered into the hole, the kelly clamped into the rotary table, and drilling was commenced by activating mud circulation and rotation of the drill stem. As drilling progressed, and cuttings brought to the surface, the kelly travelled down through the rotary table until the swivel unit reaches the table. Feed off was then stopped, rotation was slowed, and the circulation fluid allowed to continue for a short time to carry the most recent cuttings up and away from the bit and drill collars. The pump was then stopped, the kelly withdrawn and unscrewed from the drill pipe, while the latter was suspended in

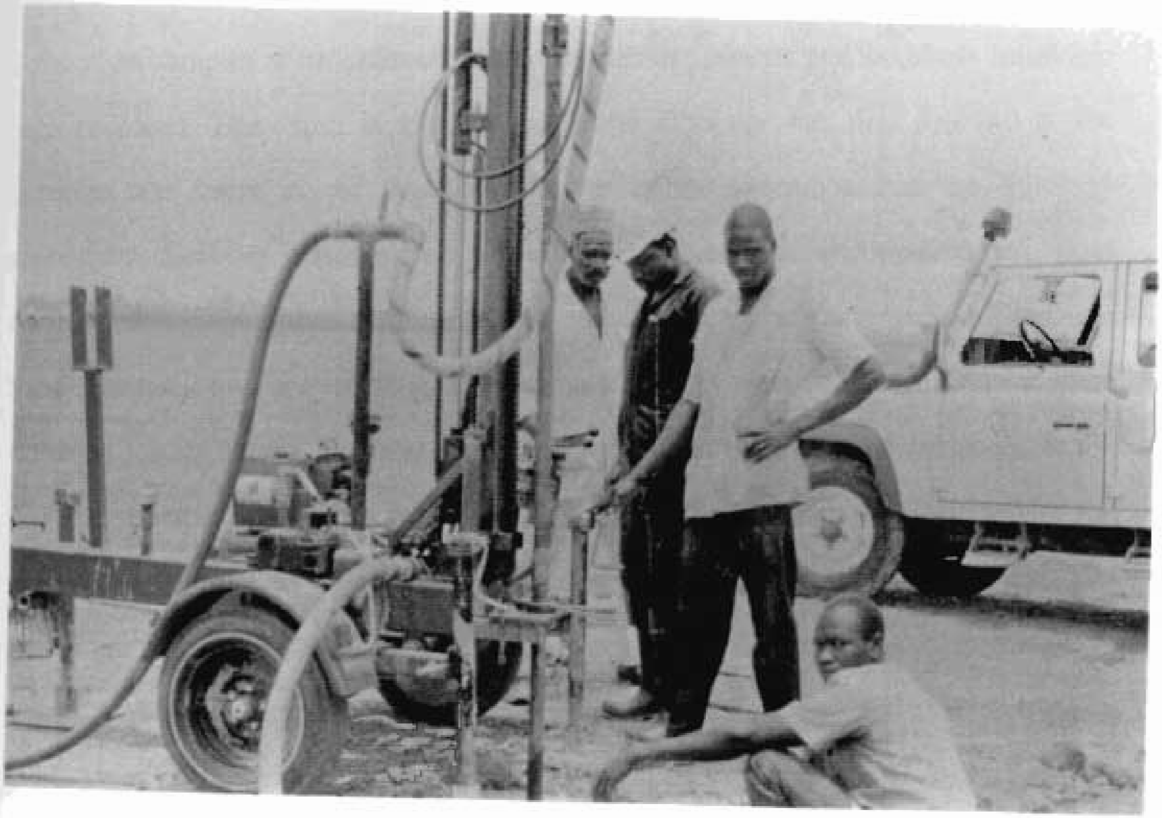


Plate 4.3. Borehole drilling in progress. The portable rotary drill rig transported to the site by a four-wheel-drive Land Rover.

the rotary table slips. Another drill pipe was added and lowered with the drill pipe column until it was at the table level, when the kelly was again attached and the circulation restarted. Rotation was engaged and finally the bit applied once more to the bottom of the hole. The above procedure was repeated until such time as the final depth (15 to 30m) was reached or the tools withdrawn to clean a blocked bit. Plate 4.3 shows drilling work in progress.

Formation samples were collected from the drilling-returns and borehole lithologic logs prepared. The depth at which a particular lithology was first detected in the cuttings was taken as the depth of the top of the formation that the lithology represents. Drilling was stopped when a new layer was encountered but fluid circulation was continued until no further cuttings were being returned. Drilling was then continued over the sampling intervals and cuttings were collected. Samples were taken at 1.50m interval by passing the drilling returns over a container placed at the edge of the drill hole. The 1.50m sampling interval corresponds to the length of the drill pipe, the depth at which drilling was stopped and new drill pipe was added. The depth control of the cuttings taken in this way was good and cross-contamination of samples was reduced. The collected samples were dried and the formation descriptions were recorded in a standard daily drilling log sheet.

The author was personally involved in all the drilling and sampling operations.

Monitoring Piezometer Network

Monitoring of alluvial groundwater was accomplished by the use of observation piezometers installed in the floodplain alluvium. The aim was to:

1) determine groundwater level changes in the alluvium and consequently to determine groundwater recharge;

2) establish river-groundwater relationships; and

3) establish groundwater flow patterns in the alluvium.

Each piezometer consists of a 50 micron, 50mm diameter 150mm long porous ceramic cup attached to a 19mm diameter rigid uPVC pipe. A porous cup design was used to avoid the piezometer being filled by fine-grained material from the aquifer. Also a smaller diameter uPVC pipe was used in order to minimise the lag time between groundwater level changes and piezometer response.

Each of the piezometers was placed in a 100mm diameter hole drilled using conventional rotary drilling method. The drilling was done with 100mm diameter drill bit (the drilling operations have been described earlier), and the aquifer was penetrated to more than 4m in each case to ensure the porous cup was not placed in a thin layer of sand. When the required depth was reached, the drill rod and bit were withdrawn, and porous cup with 6m length of pipe attached was lowered into the hole. The top of the pipe was held in position and subsequent lengths were attached until the piezometer tip rested at the bottom of the hole. The tips were placed between 11 and 16m depth from the ground surface. The levels of the ground surface and the piezometer tips are given for each location in appendix A.

The porous cup was gravel-packed by gently pouring gravel around the pipes which passed through the rising flush of mud and settled at the bottom to form an envelope around the cup.This was done to facilitate water inflow and to protect the porous cup which can otherwise be clogged with fine material reducing its life expectancy to less than 6 months (Garret and Carter, 1983).Two head pans of gravel were used in each case. The remainder of the annular space was sealed up with bentonite pellets and clay to prevent vertical ingress of water along the pipes into the hole. About 2m of bentonite was placed in the hole at the level of the confining layer (Figure 4.4) and the remainder of the hole was packed with clay which was sufficient to form a seal (Kruseman and de Ridder, 1990).

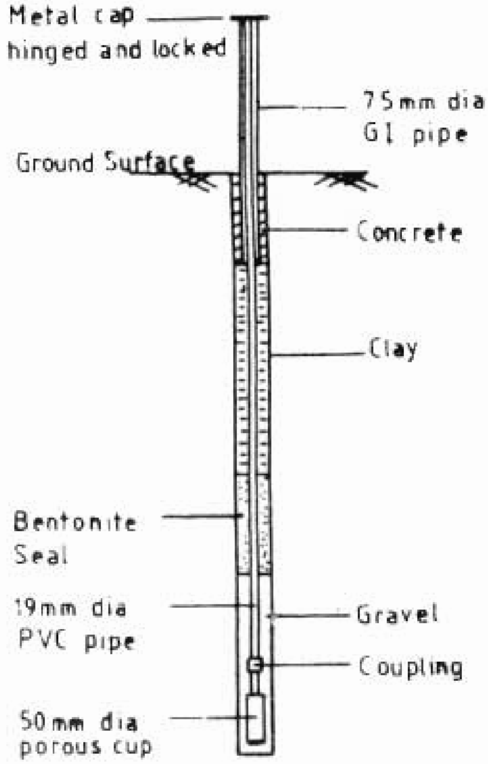


Figure 4.4. Diagram showing piezometer design and installation details.

Plate 4.4. A damaged piezometer. The protective GI sleeve removed from its foundation.

The 1.5m length of uPVC pipe projecting above the ground was protected with a sleeve of 75mm GI pipe concreted half a metre into the hole. The GI pipe was capped with a hinged plate and locked. The top end of the uPVC pipe was closed with a cork plug to prevent vertical ingress of rain and flood waters, and to avoid blockage from children. Details of piezometer assembly is shown in Figure 4.4.

Finally the ground level elevations at each piezometer were levelled into existing bench marks while piezometer locations were established by transverse surveying using a theodolite.

One piezometer per day was installed with little difficulty. However damage to installed piezometers by the local Fulani herdsmen was a serious problem. More than ten of the locks had to be replaced, and at one location an entire piezometer had to be rescued. An example of a damaged piezometer is shown in plate 4.4. 20 piezometers were installed; nine of the piezometers are located on the south side of river and the remainder are on the north side. Piezometers north of the river were named A to K while those on the south side of the river were named P1 to P9. The layout of the piezometers is shown in figure 4.2.

Piezometer water levels were monitored twice a month from 13/5/1992 to 27/6/1993 using an electric dipper. With a team of two people and a Land Rover vehicle it was possible to read all the piezometers in less than six hours in the dry season. However in the wet season access to the study site was difficult and piezometer locations could only be reached using a canoe, or walking/swimming in the flood water. Consequently water level measurements took more than twelve hours to be completed and the frequency of measurement was restricted to once a month in September and October 1992. Plate 4.5 shows water level monitoring operations in the wet season (1992).

The measured water level data were stored on a Lotus 1-2-3 type spreadsheet package for manipulation and analysis. Time-series graphs were produced for each piezometer for the duration of the monitoring period, and the 1992 Yobe River stage hydrograph

Plate 4.5. Alluvial groundwater level monitoring operation in the wet season (1992).

(a) The piezometer (H) being monitored is located at the north floodplain boundary; (b) the green vegetation is flood rice; and (c) a canoe - the means of transportation.

imposed upon it. This enabled an assessment to be made of the degree of interaction between the river and the aquifer.

River Stage

The Yobe River stage was measured using a stage board. The aim was to determine river discharge and to establish the effect of the Yobe River stage changes on the alluvial groundwater level. The existing stage board at the Gashua Bridge was used for this purpose. An additional stage board was also installed about 3km upstream of the bridge but was destroyed during flooding. River stage measurements were made at least twice a month in the dry season. However, the measurement frequency was increased in the wet season and river water levels were measured daily when the river was at its peak, and in the first month of flood recession. The measured river stages were converted to discharges using rating equations developed for the Yobe River at the Gashua Bridge (Thompson,1992).

The flood water height within the floodplain was also measured. The GI sleeves protecting the piezometer extensions were used for this purpose. The water level heights during flooding were read off directly on the sleeve using a measuring tape. Flood water heights were also measured from the marks left behind by the flood water following recession. .

Surveying

As mentioned earlier, transverse and levelling surveys respectively were carried out to define piezometer locations and to determine ground surface elevations at each piezometer location. In addition, a detailed topograhic survey was also undertaken with an attempt to prepare a topographic map for the study site. The topographic survey was carried out in the following ways. First, the coordinates of the four edges of the study site were established using a theodolite and ranging poles. This was followed by division of the study site into equal grids, i.e 250 x 250m squares.

Levelling survey followed and ground surface elevation at each intersection points was obtained.

Further, river channel cross-section dimensions were also determined using the levelling instrument and its accessories. Ten cross-sections were established across the river channel within the study site and the channel's dimensions (width and depth) were measured at each location. During each measurement the top and bottom width of the river channel was measured using both a chain and a measuring tape, while the river channel depth was read off from a levelling staff which was referenced to an existing bench mark. The river channel has a trapezoidal cross-section; the top width ranges between 41 and 54m, while the bottom width ranges between 21 and 32m. The average top and bottom widths are 46.6m and 25.4m respectively. The channel depth ranges between 4 and 5m; the average channel depth is 4.30m.

Pumping Test

Pumping tests were used to evaluate the hydraulic properties of the alluvial aquifer namely transmissivity (T) and storage (S). Two tests were conducted in July and October 1992 using a single pumping well and an observation piezometer located 4.3m away from the well.

The pumping well was constructed by the conventional rotary drilling method using 150mm diameter drill bit. The total depth of drilling was 14.5m, the bottom of the aquifer where a clay layer was encountered. The well was lined with plain and slotted uPVC casing for its entire depth, the lower 1.0m consisting of plain casing, closed by a wooden end plug, and with 5.8m of slotted pipe above this. The slotted interval was gravel-packed with chippings, while the remainder of the annulus was filled with gravel and clays. Upon completion the well was developed by continuous pumping and surging. The steps followed were: first water was pumped into the well through the drill pipes and drill mud was forced out of the hole; then the well was pumped continuously with intermittent switchoffs to create surging. This continued until clear water free of mud and fine sand was obtained.

The piezometer was constructed first by hand augering to 4.3m, and subsequently jetting using 50mm diameter GI pipe connected to a flexible hose from a centrifugal pump. The jetting pipes were reciprocated during jetting, and drilled materials were brought to the surface. Subsequently a piezometer assembly was lowered into the hole and jetted to a final depth of 8.17 m below ground level. The piezometer was made up of 8.47m of 25mm diameter uPVC pipe. Its lower 0.5m was slotted horizontally.

During each test a petrol driven pump was used for pumping, and drawdown/recovery measurements were made in both the pumped and the observation wells using an electric dipper. The drawdown was measured initially at fixed time intervals of one minute for the first 10 minutes, and the interval was increased in a logarithmic fashion for the rest of the test period. Pumping lasted for 2h 28min in July (the test being curtailed by heavy rain), and 3h in October. The maximum drawdown recorded in the observation piezometer was 0.525m, and almost full recovery was attained 3h after the pump was switched off.

Generally one of the difficulties faced while conducting pumping tests is in keeping the discharge rate constant. The rate tends to vary with increasing head. It is necessary to keep the rate constant because virtually all analytical solutions for aquifer tests require a constant flow rate to yield reliable results. A constant flow rate of 2.0l/s was maintained by connnecting a 25mm diameter GI pipe to a delivery hose which served as a discharge regulator. The flow rate was measured volumetrically using a 200l oil drum, and checked periodically. Water pumped from the well during the test was discharged into a nearby pit.

4.4 Conceptual Modelling

Following the detailed geological, geophysical and hydrological study of the Yobe River-aquifer system, a conceptual model of the interconnected Yobe River-aquifer system and the processes acting on it was constructed. This involved determining the geologic framework and groundwater regime of the river-aquifer system. The flow boundary conditions were also defined.

CHAPTER FIVE

THE GEOLOGIC FRAMEWORK OF THE YOBE BASIN ALLUVIUM

5.1. Introduction

A hydrogeologic study of an area necessarily includes the determination of the geologic framework of the system under consideration. The geologic framework of the Yobe basin within the present site was obtained from a combined use of geophysical and drill log data.

A description of the drill log data is given in section 5.2 of this chapter. Section 5.3 reports on the geoelectric data obtained and the interpretation and discussions thereof. The geoelectric and the drill log data are combined in section 5.4 and the system framework is defined. The last section (section 5.5) gives the results of a pumping test conducted within the site.

5.2 Drilling Results.

Lithology

Twenty one (21) boreholes were drilled and their lithologies recorded; the lithologies for eight of these boreholes, are shown in figure 5.1. The complete listing of the drill log data is given in appendix D. An examination of the logs indicates that:

1) the sediments comprising the alluvium are variable, and they include unconsolidated gravel, coarse and fine sands, silts and clays. Like most alluvial deposits, the sediments fine upward with the lower part of the alluvium often consisting of gravel and coarse sand while sediments comprising the top stratum are most commonly clay and other fine materials;

2) at most locations layers of sand and gravel occur in sequence; at some borehole locations a higher proportion of sand occur in the boreholes (e.g boreholes A and D),

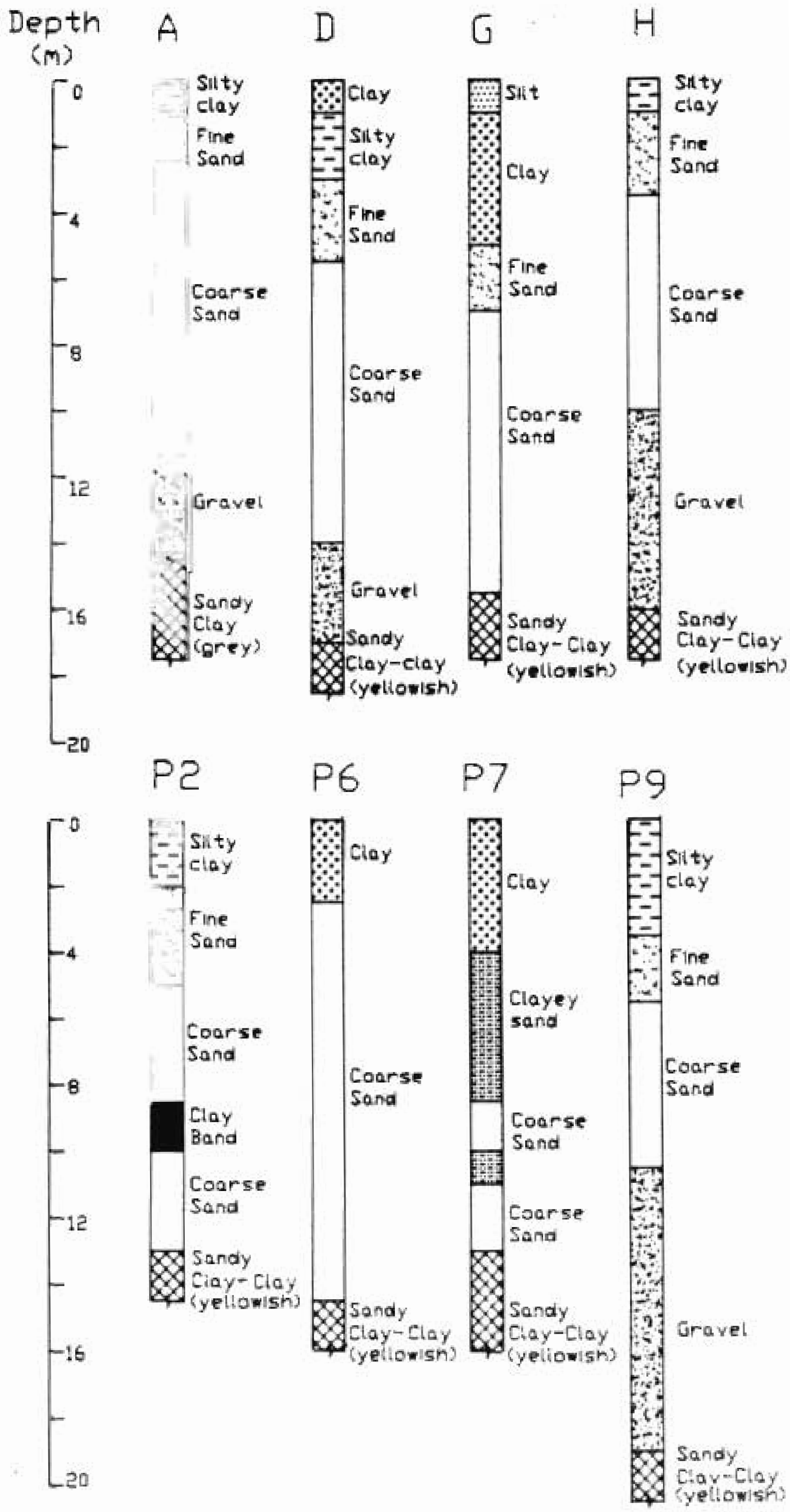


Figure 5.1. Typical borehole drill logs from the study site

at other locations the profile contains more gravel (e.g borehole P9). In most boreholes the coarse sand and gravel layers are overlain by fine sand which is in turn covered by clays and other clastic materials (figure 5.1);

3) there exists, as would be expected from most alluvial deposits, a gradual change (both laterally and vertically) from clean sand and gravel (good aquifer material) to sand and gravel intermixed with clayey material (marginal to poor aquifer material), and/or abrupt change from the former to the latter, and vice versa. For example at borehole P7 a higher proportion of the borehole depth is occupied by a sequence of clayey sand and clay;

4) clay bands of limited thickness are occasionally encountered in the aquifer, e.g in borehole P2. Depths where the drill bit encounters bands are indicated by an abrupt decrease in rate of bit penetration with frequent blockage of drill bit and consequent loss of drill fluid circulation. The size of cuttings produced from the boreholes also indicates when the drill bit enters a clay band. Cuttings produced when drilling through sand and gravel are loose coarse particles, whereas larger cuttings or 'chips' were produced when drilling through a band;

5) the thickness of the sand and gravel layer ranges between 8.50m and 19.50m with a mean of 11.70m. The top layer, comprising clays and silty clays among others, has thickness ranging between 1.0m and 5.0m and the mean layer thickness is 3.0m; and

6) the sediments described above were deposited on sandy clay/clay layer. The cuttings produced from this layer during drilling appear as slurry and the mineralization associated with it show up as a change in colour of the cuttings. The colour of the lower unit ranges between grey and buff.

5.3 Vertical Electric Sounding (VES) Results.

Accuracy of VES Data.

The sounding data are generally good with low observational errors and reasonably low offset errors. An example data set is given in table 5.1. The complete listing of VES data is given in appendix E2.

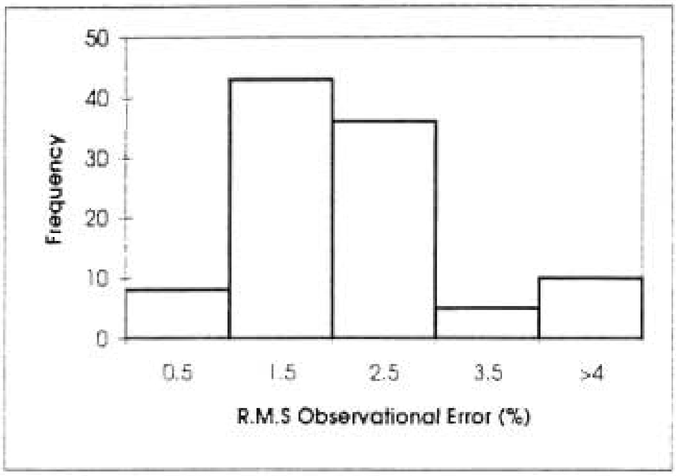
Table 5.1. Example data set: VES5

Spacing (m)	Resistances (Ohm)						Wenner Resistivity (ohm-m)	Observed error (%)	Offset difference (%)
	RA	RB	RC	RD1	RD2	RD			
0.5	3.94	0.38	3.56	2.36	4.29	3.33	10.45	-4.08	-58.04
1.0	1.97	0.09	1.88	1.26	1.57	1.41	8.88	0.60	-22.34
1.5	--	--	--	--	--	--	7.07	--	--
2.0	1.02	0.06	0.96	0.71	0.62	0.66	8.31	2.91	14.06
3.0	--	--	--	--	--	--	10.71	--	--
4.0	0.88	0.05	0.82	0.56	0.53	0.54	13.65	-1.26	5.89
6.0	--	--	--	--	--	--	19.55	--	--
8.0	0.80	0.05	0.75	0.55	0.46	0.50	25.18	-2.83	17.96
12.0	--	--	--	--	--	--	35.86	--	--
16.0	0.70	0.04	0.67	0.48	0.44	0.46	46.60	-2.42	7.98
24.0	--	--	--	--	--	--	54.19	--	--
32.0	0.37	0.07	0.30	0.28	0.29	0.29	57.91	-0.46	-4.86
48.0	--	--	--	--	--	--	71.54	--	--
64.0	0.22	0.02	0.19	0.12	0.12	0.12	48.46	0.46	0.83

RMS % observational error = 2.277; rms % offset difference = 23.748.

Figure 5.2 shows the frequency distribution of the root mean square (rms) observational and offset errors. The rms observational error is generally within or close to the 2% limit of Barker (1988) and, in this case confirms that all random errors such as instrument malfunctions, current leakage and high contact resistances at electrodes were recognised and corrected in the field. Thus the data obtained from the

VES survey in this study may be considered to be devoid of much of the observation errors.



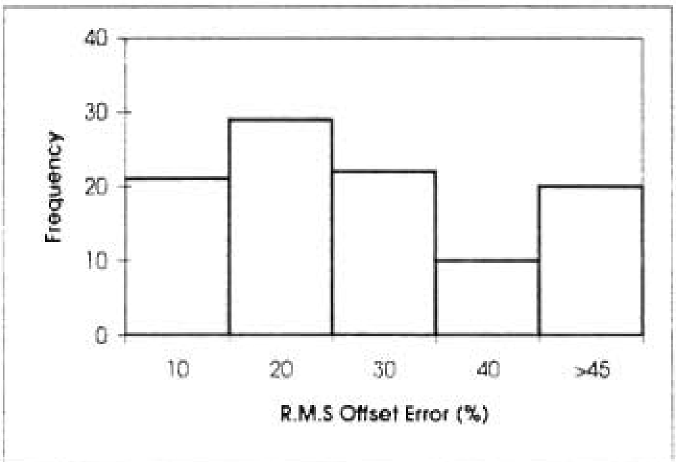


Figure 5.2. Frequency distributions of observational and offset errors

The rms offset errors show a much broader distribution than the rms observational errors. It was found that soundings with rms offset errors lower than Barker's upper limit of 25%, are those where the near surface lateral resistivity variations have caused small and normally random offset errors. These soundings produced smooth curves and allowed easy interpretations.

However about one third of the soundings exhibit higher rms offset errors (greater than 30%)- indicating the differences in geology at different points along the lines of the measuring electrodes. The deviation of field data points from theoretical curves on some of the soundings curves depict the offset errors associated with near surface layers and those arising inside the heterogeneous alluvium.

Though sounding curves with higher rms offset errors are generally required to be used with caution (Barker,1981,1988), the interpreted geoelectric parameters from the curves in this study agreed well with adjacent borehole logs and thicknesses derived from VES curves with low rms offset errors.

Typical Curves

On inspection of the geoelectric data obtained from all VES stations, the field apparent resistivity curves can be classified into two general types. Examples of the two types of curves, type A and type B, are illustrated in figures 5.3a and 5.3b by the field data for stations VES 45 and VES 26 respectively. Of the 102 field curves, 22 were classified as type A. Type A sounding curves describe qualitatively a model composed of a minimum of four layers where the layer resistivity relationship is R1>R2<R3>R4. Based on drill log information, and on field observations, the conceptual layered earth model that best fits a type A curve is:

1) a relatively thin surface layer of fine grained material (top soil, silt or clay);

2) a low resistivity layer representing the clay unit;

3) a layer of sand and gravel comprising the aquifer; and

4) a low resistivity layer representing the lower saturated clays/sandy clays.

Sixty eight (68) of the remaining field curves (measurements at 12 stations, VES 2, 11, 16, 23, 34, 38, 77, 79, 93, 95, 99 and 102 were not used because of spurious results) were classified as type B (three layer case) where R1<R2>R3. The difference

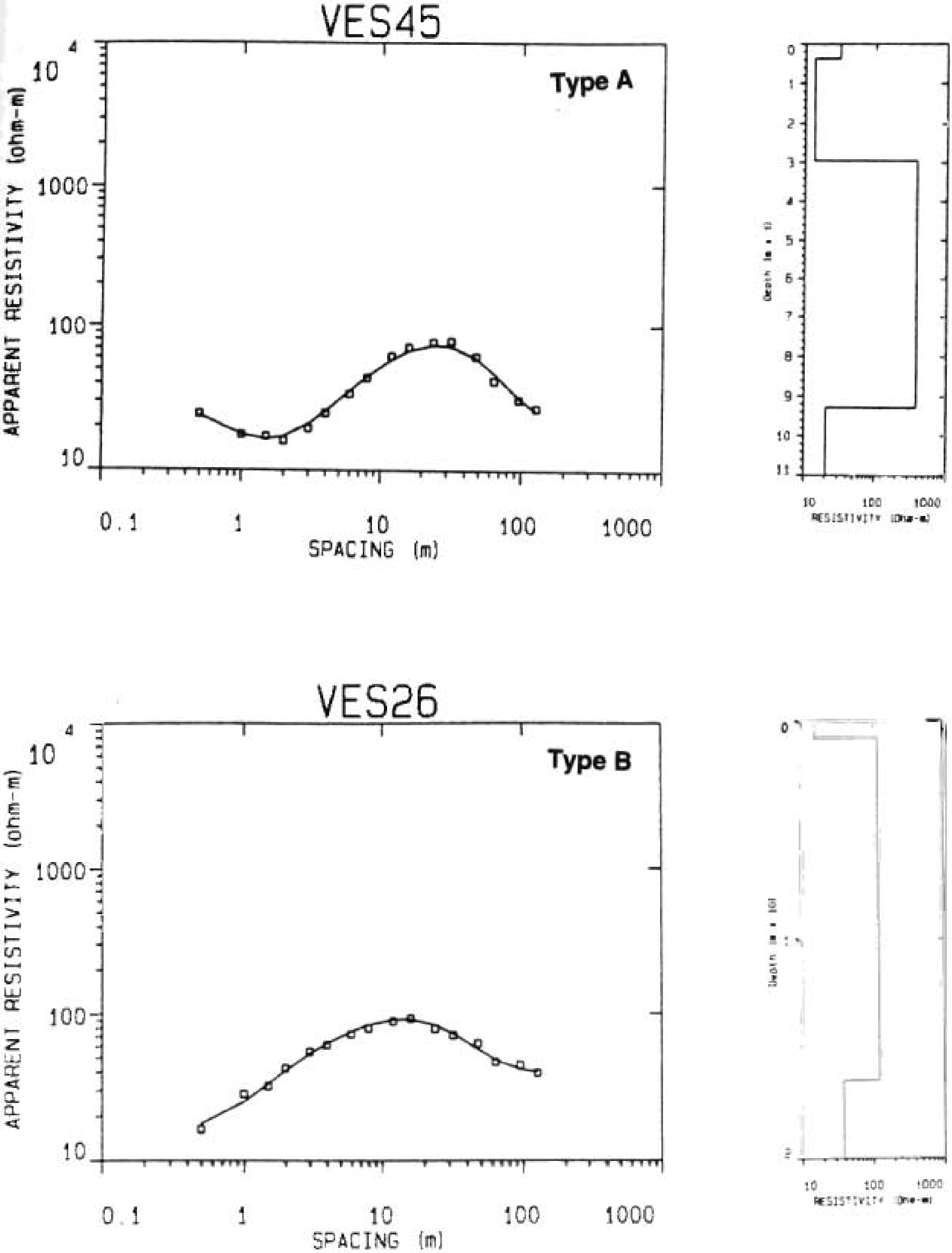


Figure 5.3. Characteristic field curves - type A and type B

between the two curves, and, hence, between the conceptual models, is the absence of the thin surface layer.

Correlation of Resistivity with Geology.

A comparison of geoelectric data with logs from nearby boreholes is shown in figure 5.4. Two salient points are illustrated by the figure. First, there is a good agreement between the locations of resistivity discontinuities and location of stratigraphic contact. The second point of interest illustrated by the figure is that the layer thicknesses derived from the VES data for the alluvial section correlate with presence of a surface clay layer and sand and gravel beneath; these units have been identified in all of the borehole logs. In general, the agreement is either exact, or within 15% ; examples are VES 31 and borehole F (separated by 125m), VES 19 and borehole K (separated by 350m), and VES 32 and borehole F (separated by 525m).

Detailed discussion of correlations at every sounding would be prohibitively long and tedious. Instead examples are discussed for a typical correlation and for situations where the comparison revealed some features of geological and geoelectric interest. Geologic profiles of four boreholes (boreholes F,G,K and P6), and twelve VES geoelectric cross-sections given in figure 5.4 were used for this purpose.

The well closest to VES 20 is K. VES 20 is about 125m north of K at about the same elevation. The top clay layer in K matches a layer of 17 ohm-m resistivity in VES 20. The interpreted thickness (0.70m) of this layer fit moderately into the clay layer in the geological profile. A moderately resistive (188 ohm-m) zone in VES 20 begins at a depth of 0.70m below the ground surface. The sand and gravel unit thickness (19.80m) derived from the interpretation of VES data is in close agreement with the thickness (19.50m) obtained for the same unit during drilling (borehole K). A conductive zone of 16 ohm-m appeared immediately under the sand and gravel unit. Drilling penetrated through the entire thickness of the sand and gravel unit into about 1-2m of the underlying clay which serves as the lower boundary of the overlying

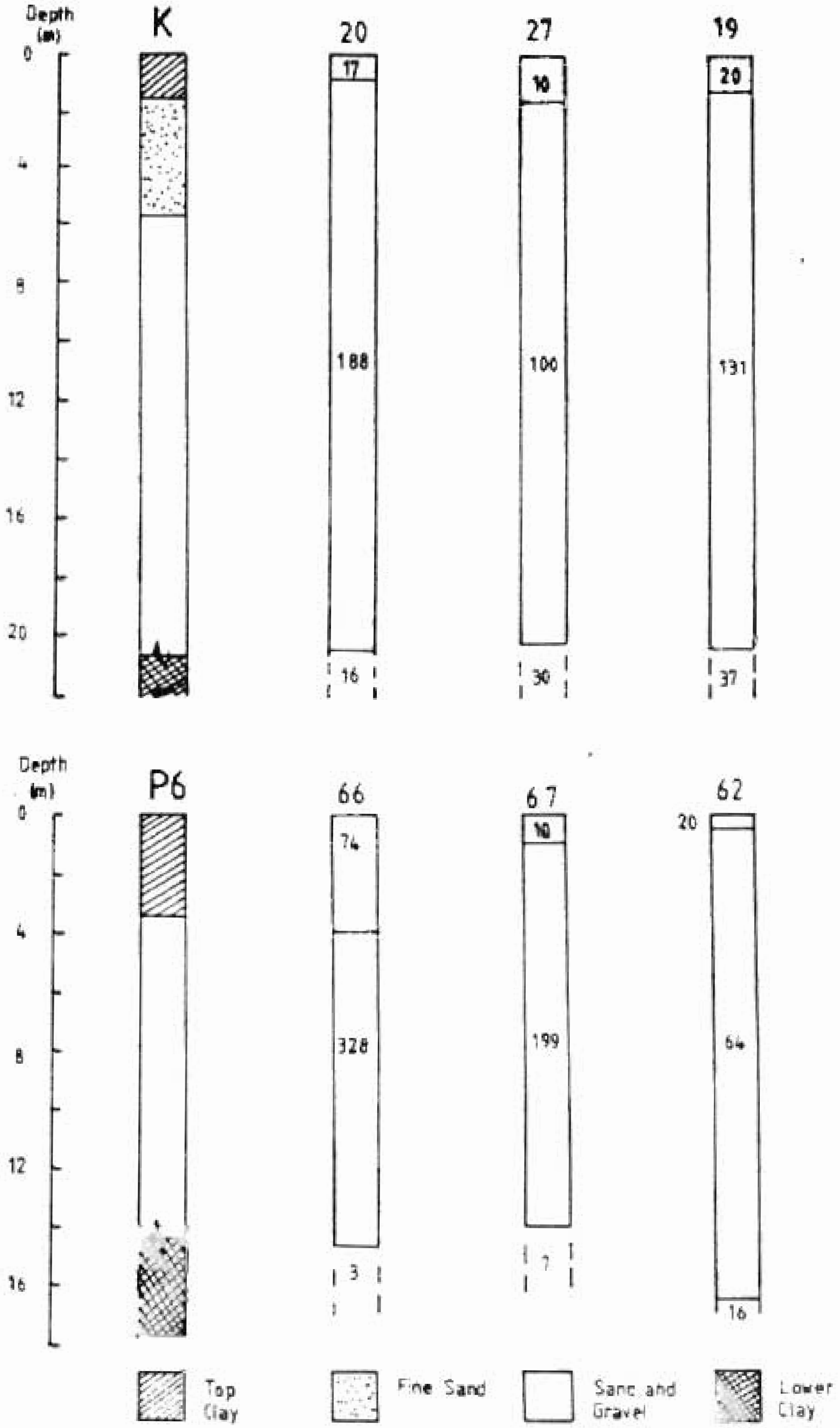


Figure 5.4. Comparison of interpreted VES field geoelectric sections and borehole logs for selected locations

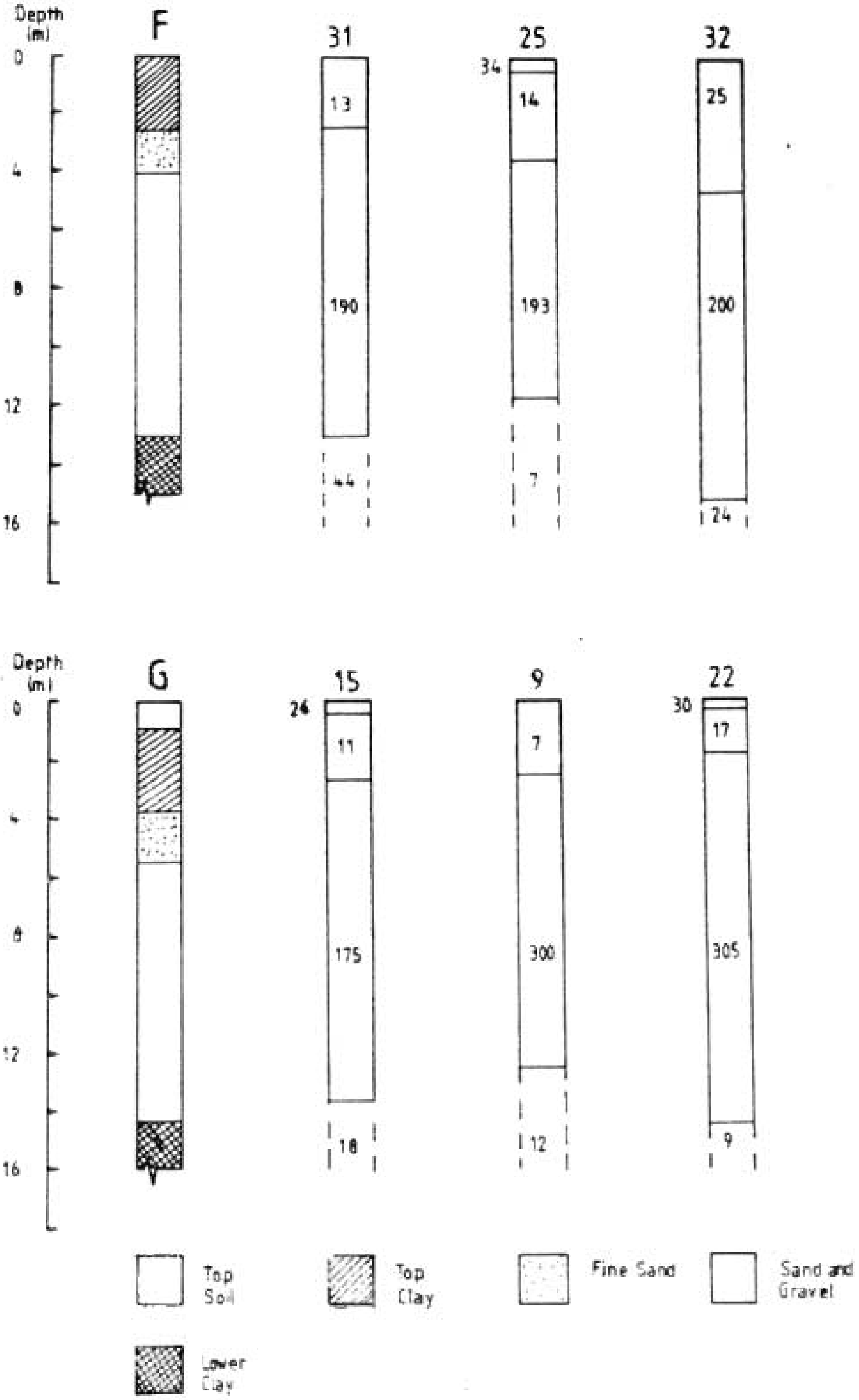


Figure 5.4. (cont'd)

sediments. Reference to figure 5.4 shows that there is a close agreement between the interpreted depth and the depth at which the underlying clay was encountered in K during drilling. It is reasonable, therefore, to assign the 16 ohm-m resistivity to this unit. The VES curves commonly show a gradual decrease in apparent resistivity as electrode spacing are increased which suggest that the lower clay unit is continuous and forms the base of the overlying sediments.

Similar results were also obtained when borehole K was compared with VES 27 and VES 19 located at distances of 425m and 350m respectively from K. In these latter cases, however, the resistivities corresponding to the sand and gravel unit are lower (less than 150 ohm-m) than that at VES 20.

The resistivity reduction becomes much more pronounced when similar comparisons were made between borehole P6 and results from VES stations 66 and 62 which are located at distances of 100m and 500m respectively from P6. As can be seen in figure 5.4 there is a close agreement between the thicknesses derived from VES 66 and the drill log at P6, and the sand and gravel unit has a resistivity of 328 ohm-m. However the result from VES 62, though shows a three layer geoelectric section, indicates that the sand and gravel unit has a resistivity of 64 ohm-m. Based upon Page (1968), and our experience here with correlation in the present study, resistivity of 64 ohm-m is low for clean sand and gravel.

The reduction in resistivity at VES 62 can be attributed to several factors. First, and most obvious, is that the sand and gravel unit pinch and swell abruptly over short distances such that it is absent at VES 62. It was found ,however, that interpretation of results from neighbouring VES stations such as VES 63 and 64 (given in Appendix E1) and drill log data do not support this proposition. Instead sand and gravel maintains considerable thickness at those locations. While thickness variations can contribute to the observed difference, the magnitude of the resistivity variation was considered large in the present study to be accounted for by this mechanism alone.

Other possible causes for the reduction of resistivities may be due to higher porosity, greater clay content or higher total dissolved solids (TDS) at VES 62. With the VES data alone, however, it is difficult to resolve the relative contributions of these factors (or other yet unknown) without independent field measurements. The effect of the presence and salinity of groundwater tends to increase the possibility of variations of the electrical resistivity and thereby extends the range of resistivity values measured. But a resistivity reduction from 328 ohm-m to 64 ohm-m would require an increase in TDS by a factor of 5 for a material of uniform porosity. This estimate assumes an Archie's law behaviour for the sediment (Ward,1965). Alternatively, adding a small amount of clay significantly lowers the resistivity. Wyllie (1963) points out that the extent to which the presence of clay influences rock resistivity depends upon the quantity and the chemical nature of this material and on the manner of its occurence. He states that in general, clay act as a separate conducting path additional to that provided by the solution in the rock pores. For the same porosity and pore water resistivity, clay saturated with fresh water will, therefore, have a lower resistivity than a medium or coarse-grained formation .

Although all of the above factors may be contributing to the resistivity reduction, it is plausible to attribute a larger part of the resistivity reduction at VES 62, and indeed at other locations within the study site, to increased clay content considering the environment under which the sediments were deposited. It is the nature of alluvial deposits to be nonuniformly layered and heterogeneous in composition, because of the changing river discharges over the period of alluvial deposition. The sand and gravel unit in borehole P7 about 150m away from VES 62 has a higher clay content than that at borehole P6.

The result from VES 100 shows a three layer curve producing geoelectric parameters which fit well into the geological profile in the vicinity (figure 5.5). However when compared closely with borehole log P2 (at about the same location), the effect of electrical averaging became apparent. The six layer geological section (P2) is represented by an equivalent three layer geoelectrical profile; the electrical properties of the thin (1.5m) clay band is embedded within the falling limb of the VES curve.

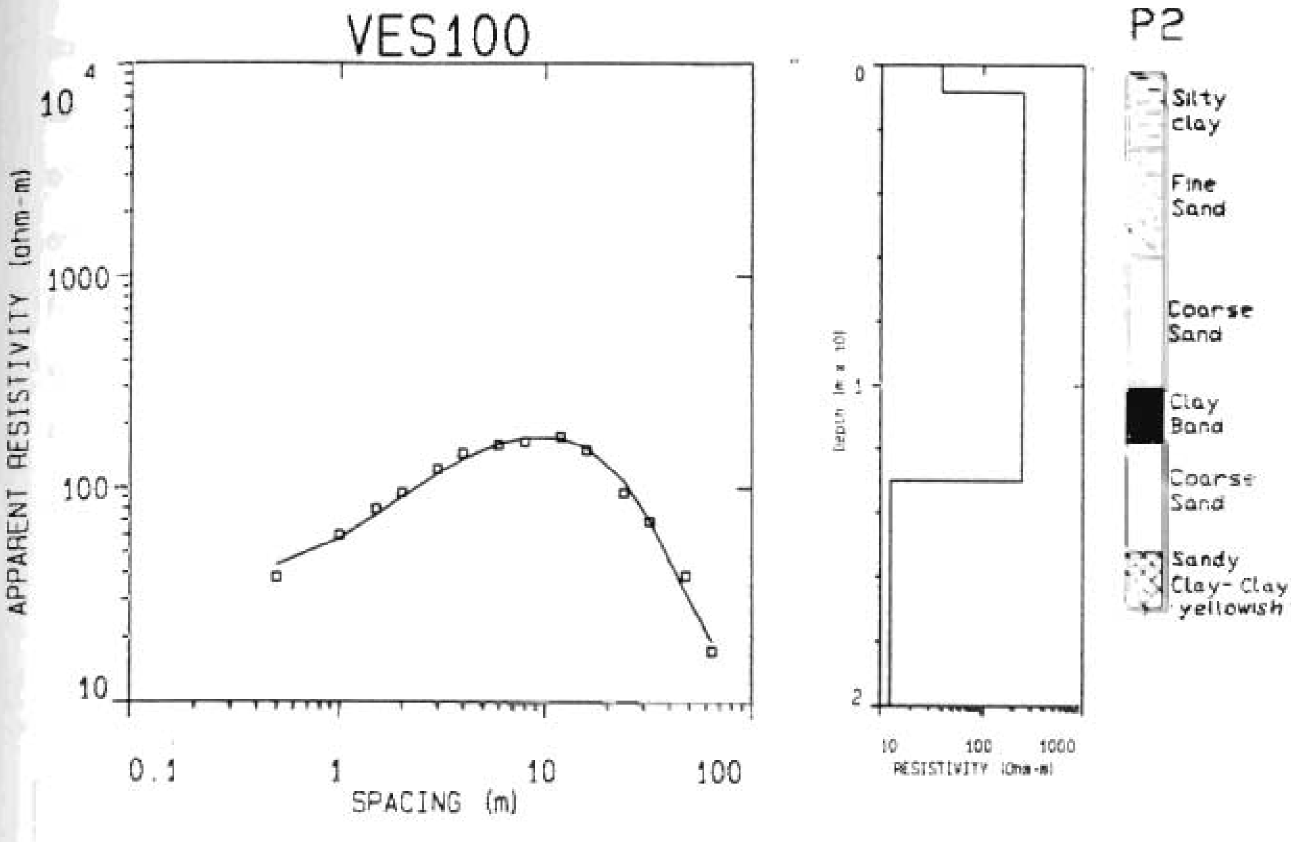


Figure 5.5. Vertical electric sounding curve near borehole P2

Suppression occurs because the thin clay band is overlain and underlain by moderately resistive layers respectively. None of the soundings made within the study site show a decrease in resistivity through these units despite the fact that thin clay bands are known to intercalate the permeable sand and gravel. The thinness of these units in comparison to the adjacent layers- the concept of "effective" relative thickness (Flathe,1955; Keller and Frischrecht,1966), and their typically lenticular and discontinuous nature, are also thought to account for this lack of resolution. For example the thin clay band encountered in P2 was not found in boreholes P1 and P3 which are located at 125m and 250m (along the same line) respectively from P2.

The effect of variations of layer properties within the aquifer was studied. Substantial changes of equivalent resistivity and thickness of the thin clay band had little or no effect on the theoretical curve shape. It was generally possible to fit the geoelectric curve to the geologic section, but thin layers within the aquifer section could not be identified from the sounding curve without the aid of lithologic data. Thus correlation of such individual clay unit was deemed too speculative in the present study because of the widely spaced borehole informations, subjective geologic drill log descriptions (e.g 'clayey sand' and 'sandy clay'), typically uncertain geophysical responses of these units, and complex sequential arrangement of these fine-grained fluvial strata. Consequently thicknesses derived for the sand and gravel unit from the interpretations of VES data include the thickness of these thin clay units.

Table 5.2 shows how thicknesses obtained for the sand and gravel unit from geoelectric interpretations fit into the geological profile based on drill log information.

Table 5.2. Comparison of geophysical interpretation with borehole data

Borehole	VES Station	VES distance from borehole (m)	Sand and gravel thickness (m) estimated from		Estimation difference (%)
			Borehole	VES data	
E	32	525	10.50	10.50	0.00
F	31	125	10.50	10.50	0.00
F	33	800	10.50	10.70	1.90
K	27	425	19.50	18.55	4.87
P2	100	10	10.50	12.15	15.71
P6	66	100	11.50	10.50	8.70

A detailed examination of the data summarised in table 5.2 and figure 5.4 shows two specific ways in which the geoelectric results differ from the drilling reports. In a few examples, the geoelectric results attribute a change in resistivity to exact depths at which change in lithology occurs. Examples are the agreement between borehole E and VES 32 separated by 525m, and borehole F and VES 31 separated by 125m. In these cases the thicknesses estimated for the sand and gravel unit are in a very close agreement with those obtained from drilling. Note, however, even in these examples the geoelectric results do not recognise the presence of the fine sand unit indicated by the drilling reports; the VES measurement considers the fine sand unit as an integral part of the lower coarse sand and gravel unit. The most frequent examples correspond to a difference in thickness estimates made for the sand and gravel unit from the drilling reports and geoelectric data. Examples are borehole K and VES 27 separated by 425m, borehole P2 and VES 100 separated by 10m, among others. The interpretations of resistivity data yielded sand and gravel unit thicknesses which are generally within 10% of actual borehole values.

There are several reasons why geophysical surveys do not produce results which are of the same accuracy as those obtained from drilling. First, it should be remembered that geophysical surveys are an indirect investigation technique. Unlike drilling which makes direct contact with a geological horizon and should, therefore, be able to determine its depth accurately, the geophysical technique attempts the same indirectly from measurements made at the ground surface. It is also likely that some of this disagreement results from the practice in resistivity interpretation of plotting depths vertically below the sounding location without corrections for the non-uniformity and heterogeneity of the alluvial deposits. Frequent interbedding of clay and sand deposits, as observed in borehole P2 , often lead to macroanisotropy and can cause discrepancy to exists between estimates made using resistivity survey and those obtained from drilling (Frohlich,1974). Only when every subsurface layer (including thin clay layers) has been correctly identified in the VES field curve, and after all corrections are made for the nonuniformity of the sediments, are the estimated thicknesses likely to be correct. Nevertheless, if allowance is made for the difficulty in identifying depths accurately during drilling, a precision in resistivity depth sounding of commonly less than 10% obtained in the present study was considered adequate for aquifer delineation.

It is generally hard to construct accurate logs of well in alluvial deposits especially using rotary drilling method. Problems that arise using this method are:

1) the recognition of clean sand and gravel units from sand and gravel units containing high clay content;

2) the mixing and homogenisation of sediments from different depth by drilling mud; and

3) the possibility that silt and clay may be washed away before reaching the surface.

Thus the differences recorded in this survey may be considered as just being outside geophysical precision.

The correlations presented above, apart from providing useful informations about the sediments encountered in the Yobe basin, also highlight some of the inherent limitations of the electrical method in aquifer delineation, i.e, the uncertainty in geoelectrical parameters determination for thin layers of geological profiles due electrical equivalence (Koefoed,1979), and the difficulty in distinguishing between clay content, porosity, and salinity effects on the sediments' resistitvity values.

Sediments' Resistivity Values

In general, attempts to match conductive zones with clays and silty/sandy clays, and moderately resistive units with sand and gravel produced satisfactory results. From the correlations, based on Page(1968) (table 5.3), and the ranges of resistivity values (figure 5.6) obtained for the sediments in the upper section of the Yobe valley (Hazell et al,1988), range of resistivity values can be assigned to different units of the alluvium:

1) resistivities between 1 and 30 ohm-m represent damp to moist clay and silty clay;

2) resistivities between 30 and 60 ohm-m represent silty clay or clay, dry to damp; and

3) resistivities between 60 and 100 ohm-m represent sandy/silty clay; and

4) resistivities between 100 and 450 ohm-m represent sand and gravel.

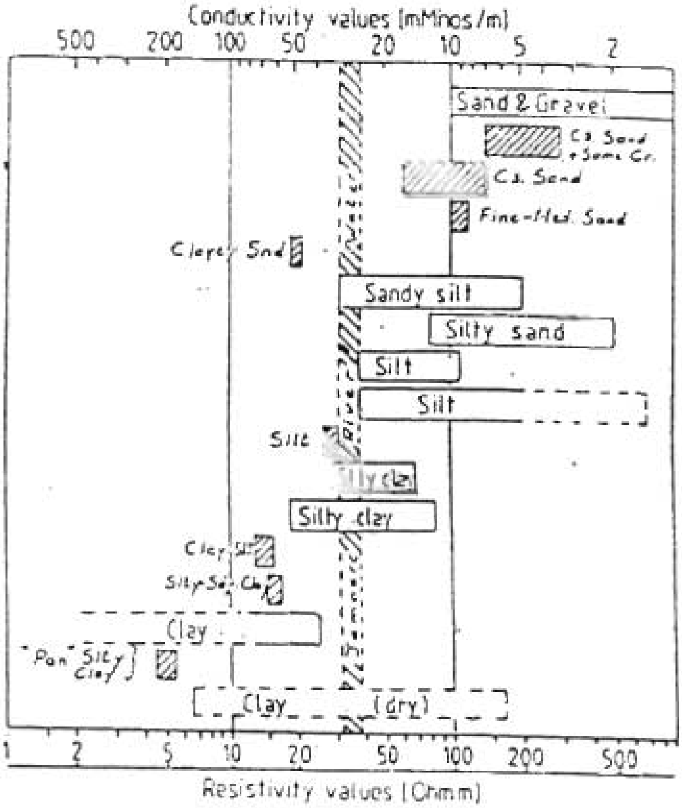


Figure 5.6. Resistivity values for alluvial sediments and water, Jama'are River (hatched) and general (open) (After Hazell et al., 1988)

Table 5.3. Resistivity data and soil types (After Page, 1968)

True resistivity range of values (ohm-m)	Soil type
1-30	Silt or clay, damp to moist
30-75	Silt or clay, dry to damp top soil; gravelly clays; clayey gravels
75-150	Clayey gravels; silty sand and gravel; saturated clean sand and gravel; interlayered sand, gravel and silt
150-1,000	Silty sand and gravel; saturated clean sand and gravel; dry, clean sand and gravel

These correlations were carried out for every sounding, and the resulting interpretations were used to define the hydrogeologic characteristics of the alluvium within the study site.

Results from the quantitative analysis of the geoelectric data are listed in table 5.4 which lists the results for selected twenty (20) VES stations according to the interpreted correlation between the geoelectric section and stratigraphic section. The complete listing of strata thicknesses and resistivities is given in appendix E1. Descriptive statistics for strata thicknesses and resistivities is also given in table 5.5.

From direct field observation and the correlations, the top layer is composed of soils, sandy silts and clays, or clays. The thickness of this layer is usually less than 1m, and the resistivity of the layer varies within the study site.

The second layer is the clay unit, which is represented by low resistivity values in most of the field curves. Although the average layer resistivity is 30.11 ohm-m, resistivities from sixty three (63) VES stations are less than 30 ohm-m which suggests that the unit is composed primarily of wet clay or at least saturated very fine-grained material. Values greater than 30 ohm-m probably indicate a mixture of coarse-grained

Table 5.4. Layer thicknesses and resistivities from automatic interpretation procedure

VES Station	Surface layer		Clay unit		Sand and gravel		Underlying Clay R (Ohm-m)	Interpretation closure error %
	H (M)	R (Ohm-m)	H (M)	R (Ohm-m)	H (M)	R (Ohm-m)		
3	0.27	47.83	1.62	12.65	10.07	103.10	33.46	8.11
5	0.66	11.00	1.58	4.84	9.32	268.80	35.12	8.00
6	0.34	72.18	2.68	8.90	12.88	293.00	16.90	5.18
7	0.34	18.00	1.66	8.50	10.00	350.00	25.00	5.12
8	0.73	22.91	2.58	11.30	9.94	236.40	52.10	7.72
15*	0.32	25.50	2.18	11.00	11.50	175.00	18.00	6.58
21	0.40	52.00	4.10	14.00	10.50	220.00	25.00	8.98
4			0.43	76.80	17.77	313.60	30.19	8.74
9*			2.50	7.00	10.00	300.00	11.50	8.40
10*			1.87	12.59	18.74	254.40	9.47	7.89
12			3.14	13.55	8.43	275.20	1.77	7.83
14	Absent		0.26	16.00	13.74	180.00	34.50	7.49
17			0.70	29.23	20.30	240.00	78.00	9.78
19*			1.25	20.00	19.00	131.00	37.00	8.96
20*			0.67	17.32	19.80	188.80	15.82	9.16
28*			9.00	64.00	19.00	100.00	3.00	9.92
31*			2.50	12.50	10.50	190.00	44.00	7.58
32*			4.50	25.00	10.50	200.00	24.00	8.52
43			0.14	66.00	13.06	187.00	34.00	9.82
46			1.68	13.81	6.58	339.60	4.19	9.88
58			0.51	109.00	12.07	457.30	1.84	4.81

H = Layer Thickness; R = Layer Resistivity, and
* indictes computations constrained by lithologic data from drill logs.

Table 5.5. Descriptive statistics for strata thickness and resistivity

Statistical parameter	Surface layer		Clay unit			Sand and gravel			Underlying clay Resistivity (Ohm-M)
	Thickness from VES data (m)	Resistivity (Ohm-M)	Thickness (m) from		Resistivity (Ohm-M)	Thickness (m) from		Resistivity (Ohm-M)	
			Well data	VES data		Well data	VES data		
No of data	22.00	22.00	20.00	87.00	87.00	20.00	87.00	87.00	87.00
Mean	0.63	51.47	3.03	1.73	30.11	11.70	12.62	253.33	17.89
Median	0.40	29.72	3.25	1.25	18.79	10.75	11.70	236.10	13.00
STDEV	0.49	62.57	1.19	1.54	25.34	2.66	3.46	86.80	13.77
Minimum	0.23	11.00	1.00	0.14	4.29	8.50	6.58	100.00	1.50
Maximum	1.99*	291.40**	5.00	9.00	109.00	19.50	22.15	457.00	78.00

* Thicknesses greater than 1.0m were only encountered at four VES stations.

**Resistivities greater than 100 Ohm-m were only encountered at two VES stations.

clastic and very fine-grained sediments. This mixed sediment is noted in some of the drill logs as silty/sandy clay. Thickness of the clay unit range between 0.14m and 9.0m with an average value of 1.73m (table 5.5). The frequency of occurence of the clay thickness is given in figure 5.7. The histograms shown in figure 5.7 indicate skewed distributions of clay thickness obtained from the VES and well data. A higher proportion (about 75%) of the clay thicknesses derived from the VES data are less than 2m thick. In contrast majority of the clay thicknesses obtained from well data are commonly thicker than 3m. The skewness between the two distributions could be due to the sample size used; only 20 borehole logs were used in the construction of the well histogram which is small in comparison with the larger (87) sample size used for the VES histogram. The smaller the sample size the less representative would be the statistical parameters obtained for the clay unit.

The third layer is composed of permeable sand and gravel. This layer is characterised, for the most part, by an intermediate layer resistivity (200 to 300 ohm-m), although the range of values is relatively large (100 to 457 ohm-m). High resistivities (e.g over 250 ohm-m) indicate a fairly clean (clay free) sand and gravel, compared to that indicated by lower resistivities where the sand and gravel is intermixed with clay or other fine-grained materials. An average resistivity for the sand and gravel is 253.33 ohm-m. The range and mean value obtained for this unit do not include results from three VES stations (VES 61,62 and 80) because the resistivity values obtained for the sand and gravel unit at these stations were less than 100 ohm-m. Thickness of this layer ranged between 6.58m and 22.15m with an average of 12.62m (table 5.5). The distribution of the sand and gravel thickness is also shown in figure 5.7. It can be seen that both the VES and well data produced similar histograms for this unit. The similarity of the two histograms indicates that the VES survey was able to estimate the sand and gravel thicknesses with almost the same precision as those obtained from drilling. In both cases a higher proportion of the thicknesses obtained for the sand and gravel unit falls between 10m and 12m (figure 5.7). The compatibility of the two

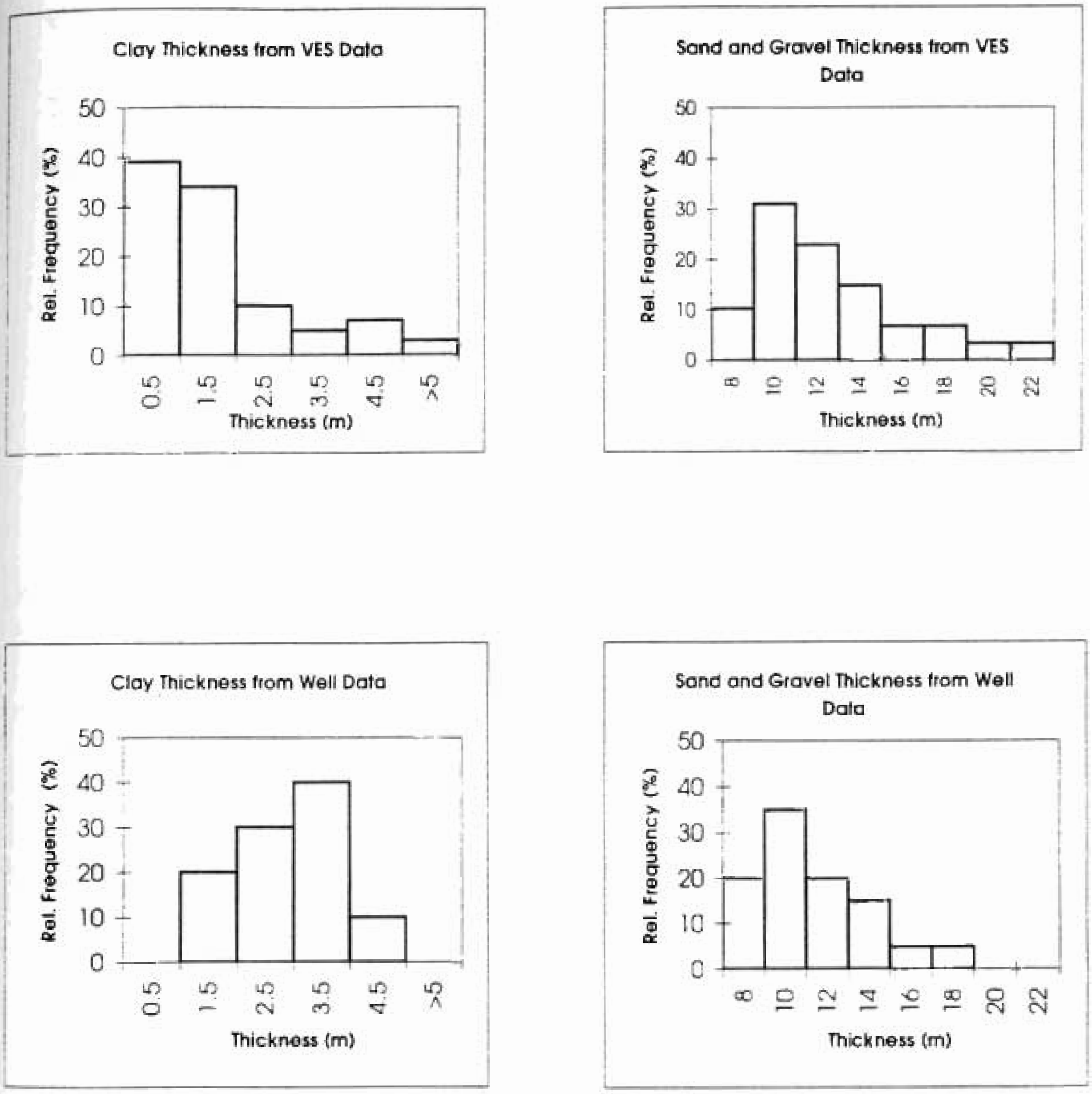


Figure 5.7. Histograms of well and VES data

results is practically important because drilling is usually very expensive and time consuming and complementing it with a fast and relatively inexpensive geoelectric survey would be desirable for delineating alluvial aquifers in the region.

The sand and gravel overlie the lower saturated clay/sandy clay. The resistivity of the underlying clay range between 1.50 ohm-m and 78.0 ohm-m, with an average value of 17.89 ohm-m. Its thickness is not resolved by the geoelectric data, and the boreholes penetrate it only partially. At one location (borehole P5), however, it is seen to has a thickness of 12m. To determine the thickness of this layer within the study site would have required drilling most of the boreholes (because there is uncertainty about the extent of this unit) to intercept the lower boundary of this unit, a technically difficult and expensive operation within the scope of the present study. Nevertheless, sufficient validation was obtained during the course of drilling to confirm the existence of this unit at 21 borehole locations and made it possible to extrapolate, using geoelectric survey, with reasonable confidence into locations with no geological data.

5.4. Sedimentary Stratigraphy and Lithofacies

Geologic and geoelectric data obtained during the course of the present study can be combined at various scales to help determine the types and dispositions of the sediments within the Yobe basin. Particular attention has been focussed on the size, shape and arrangement of the component depositional and erosional elements of the strata, which together comprise the sedimentary stratigraphy. At the study site, division of the upper 20-30m of the Yobe basin sediments into two laterally continuous, lithostratigraphic units, the Holocene alluvium and Plio-Pleistocene(?) Chad Formation, permits visualization of the large-scale stratigraphy of the basin. However, because zones of similar textural and hydraulic character are seen in the upper section of the Yobe to cross the contact between these lithostratigraphic units (Hazell et al, 1988), their values as hydrostratigraphic guides is limited. The distribution of sediments within the Yobe basin may be conceptually simple but stratigraphically and lithologically complex to be represented in such a simplified way using limited site specific data. To reconcile this difficulty, the alluvium strata were considered as smaller, more hydrologically significant, three dimensional bodies of similar textural character term lithofacies.

Strata of the Yobe alluvium in the study site were differentiated into four lithofacies based primarily on texture and natural resistivity response:

1) surficial deposits of aeolian sands;

2) top clay layer;

3) sand and gravel layer; and

4) underlying clay layer.

The general lack of intact samples prohibited inclusion of data regarding primary sedimentary structures and quantitative grain size distribution in lithofacies definition. An interpretation of the lithofacies data is presented in maps in figures 5.8 and 5.10 .

Lithofacies cross sections supplement the data from lithofacies distribution maps and provide a more detailed perspective from which to view sedimentary stratigraphy. The principal, approximately north to south-oriented, cross-sectional transects constructed (lines A-A',B-B'(figures 5.8, 5.9 and 5.10)) reflect availability of borehole geologic data along lines nearly parallel to the inferred hydraulic gradient. A single, approximately east to west-oriented, transect (line C-C') provides a three-dimensional perspective. Maps and cross sections shown in figures 5.8 to 5.10 depict the vertical and horizontal distribution of permeable and relatively impermeable units and display almost the full stratigraphical sequence within the study site.

Surficial Deposits

Surficial deposits comprise the aeolian sands of the uplands and are above the alluvial groundwater level; therefore, they are assumed not to play a significant role in the hydraulics of the subsurface system.

The Upper Clay Unit

This unit comprises clays, and sandy and silty clays. The unit has a lithology similar to those found upstream but is generally thicker and contains higher clay content. The clays are cracked at most locations and occassionally intercalated by small lenses and/or thin layers of silt and fine sand.

Interpretation of VES data indicates that the clay unit, originally identified in drill logs, is a relatively thick and extensive layer. The isopach (thickness) map shown in figure 5.8 was constructed using geophysical and drill log data. As shown by the contours, a clay layer of more than 1 metre thickness covers most part of the floodplain surface within the study site. The clay unit generally thins towards the river in the central portion of the floodplain and appears to be somewhat lenticular in shape. The clay unit thickness is less than a metre at the centre and increases rapidly to more than 4 metres to the east over a short distance. The thick clay unit tapers towards the eastern fringe of the study site where its thickness drops to 1 metre. The clay unit also increases in thickness west of the central thin clay region and decreases to the north and southwest. The clay unit picks up thickness near the north-west corner of the study site and maintains 4 metres thickness there. Also from the contours shown, the clay unit appears to be in direct contact with the upland margins and therefore does not pinch out, but rather retains a considerable thickness especially along the southern margin of the floodplain. The clay unit is 6 metres and 1 metre thick near the southern and the northern margins of the floodplain respectively.

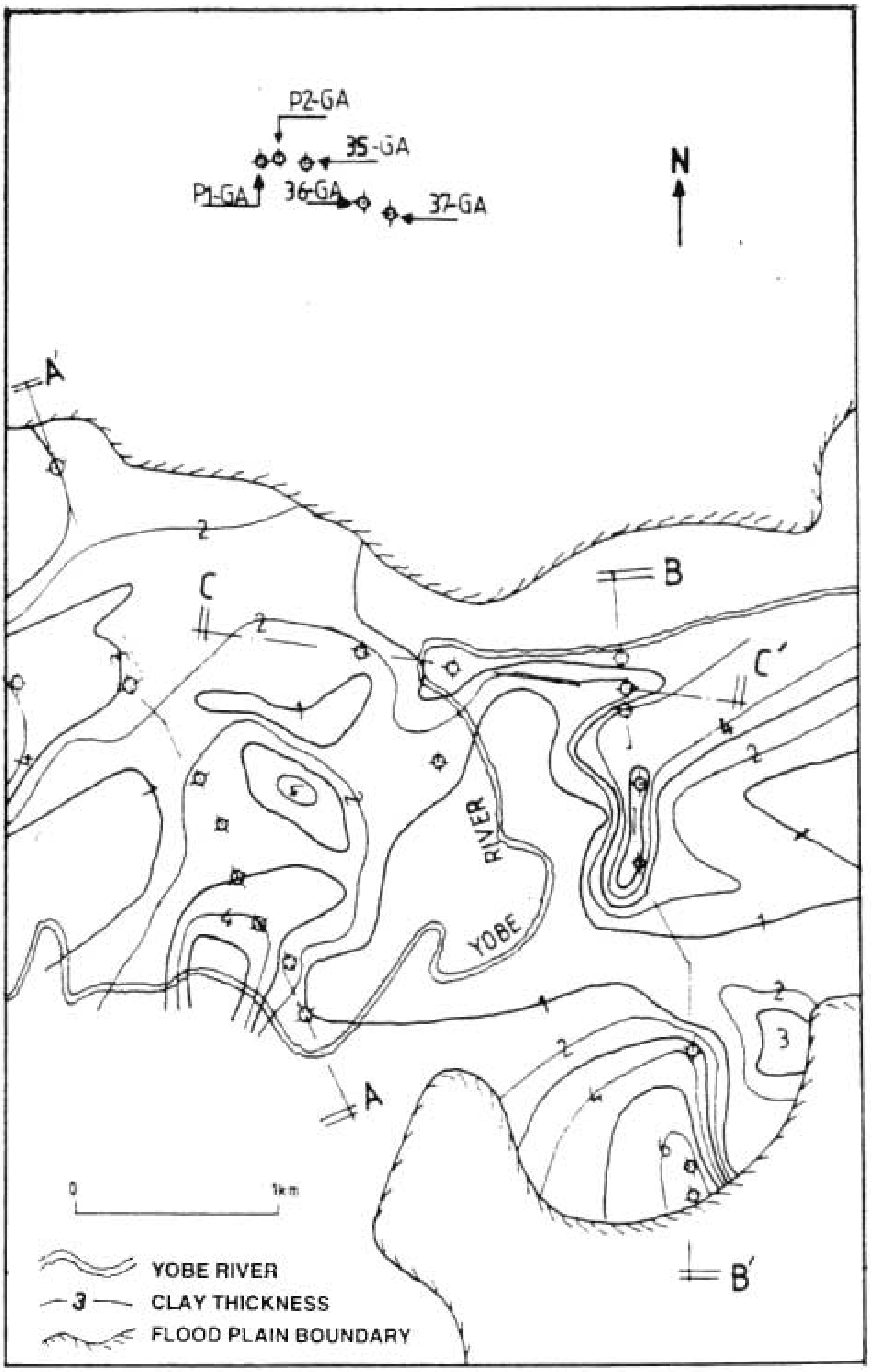


Figure 5.8. Isopach map of the upper clay unit. Both geophysical and borehole logs data were used in the construction of the map.

Because this unit is extensive and composed of media of low permeability, it forms a cover layer over most of the alluvial sand and gravel and at some locations confined the aquifer. The effect of this is to preclude the general assumption of unconfined conditions made in evaluating the alluvial aquifers in the region. This approach need modification to account for the probable differences in aquifer behaviour and effects on the water budget under confined and unconfined conditions. Some of the main factors which are intimately related to the stratigraphy which need consideration are:

1) distribution of recharge;

2) vertical distribution of hydraulic heads; and

3) boundary conditions.

Sand and Gravel Layer

Beneath the upper clay unit deposits of sand and gravel occur and overlie the underlying clay unit. The sands and gravels, as has been described in section 5.2, are at some locations clean (clay free), but ocassionally intermixed with clay and other fine materials. Clay bands and lenses are also common which seldom exceed 2m in thickness. Each individual band appears to have limited lateral extent, although more than one unit can be encountered in a given borehole log. For example the clay band in P2 was not encountered in boreholes P1 and P3 which are 125m and 250m respectively away along the same line (figure 5.9).

An isopach (thickness) map of saturated sands and gravels is shown in figure 5.10. Data used in the construction of the map were derived from the interpretation of VES field curves and drill log data. Because the floodplain is virtually flat, with the only surface relief exhibited by shallow depressions, the saturated thickness of the sands and gravels is a function of the following:

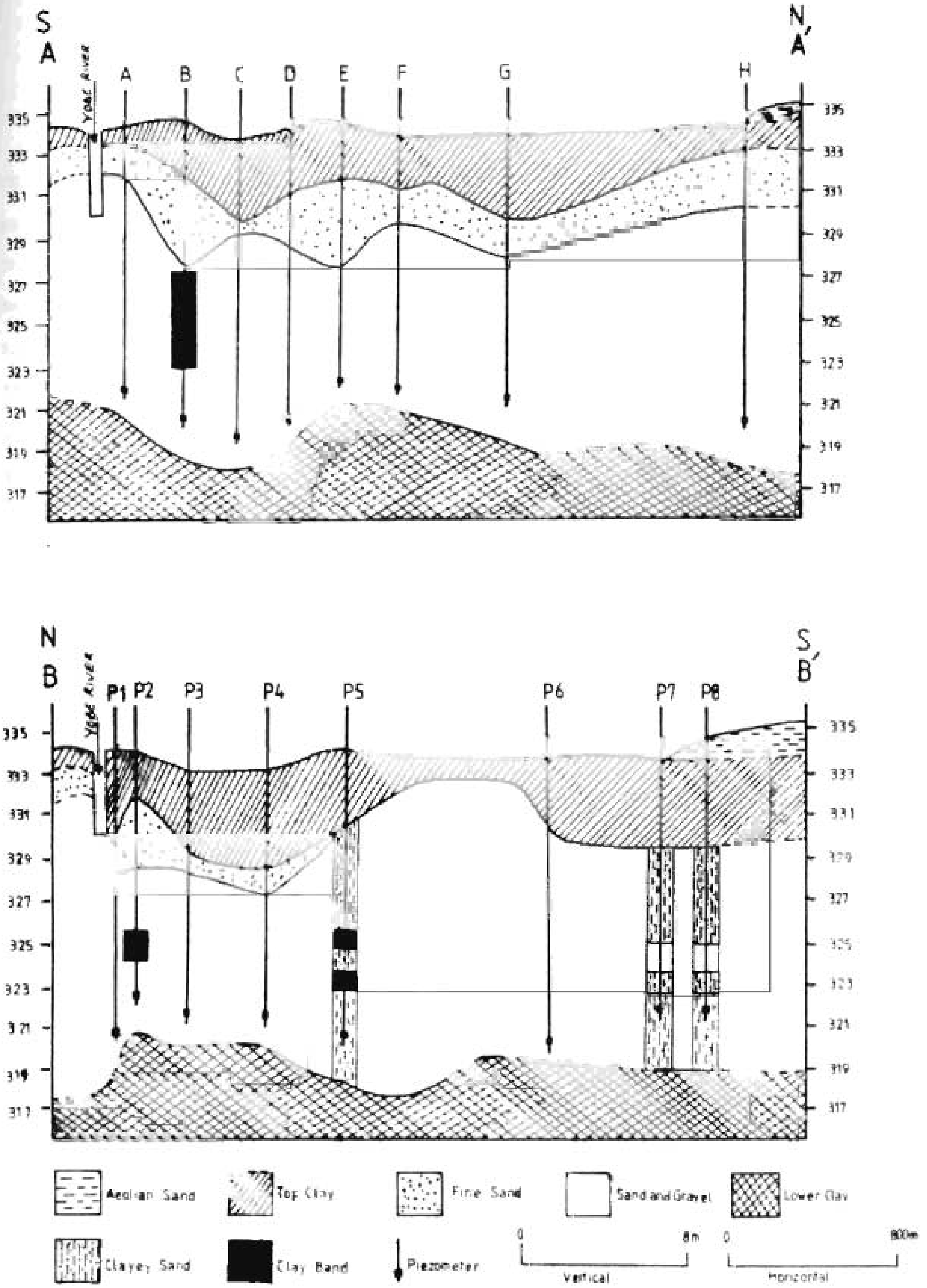


Figure 5.9. Hydrogeologic cross-sections through selected locations across the valley in the study site

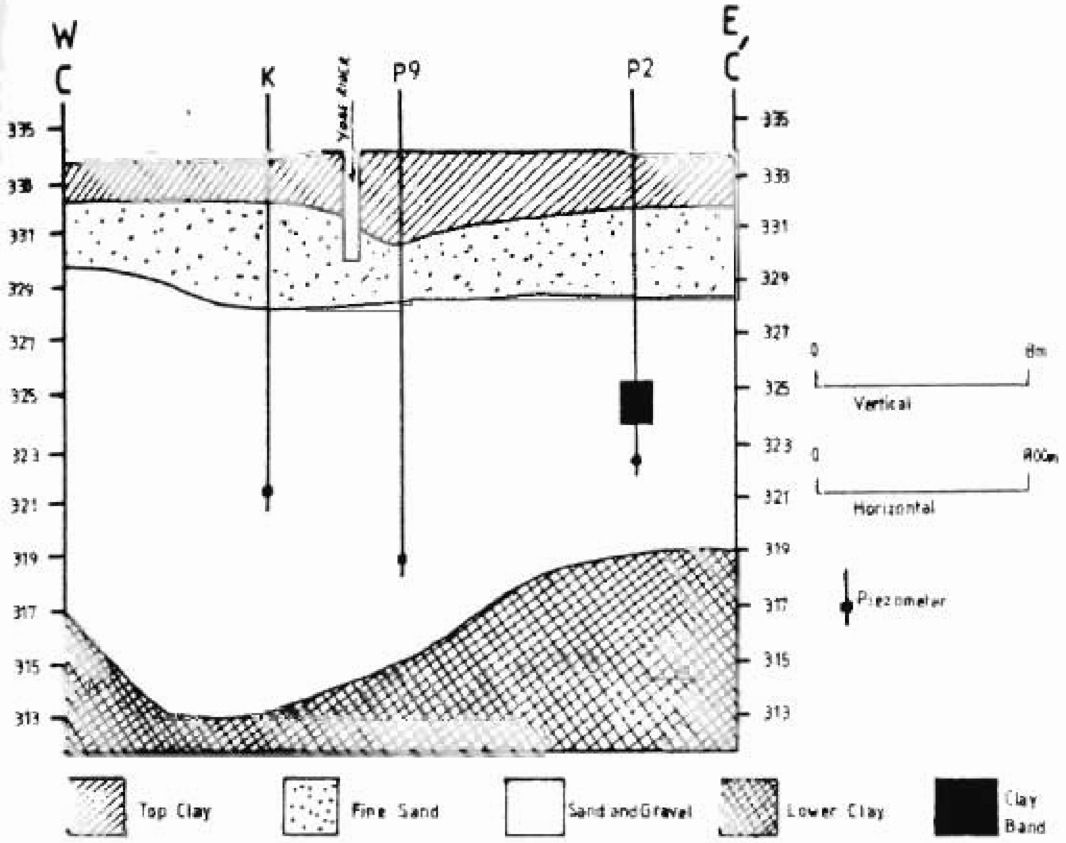


Figure 5.9. (cont'd)

1) the elevation of the underlying clay surface;

2) the thickness of the overlying clay unit; and

3) the groundwater level (located 2.5 to 3.5m below the ground surface).

An interpretation of map (figure 5.10) and cross-sections (figure 5.9) indicates that a sand and gravel layer of variable thickness occur throughout the study site. The thickness of this layer commonly exceeds 10m and thicker deposits of this unit are associated with topographic lows (troughs) on the surface of the underlying clay; and at some locations the thickness of this unit is reduced by the ridges on the surface of the lower clay. The thickness of the sand and gravel layer is also reduced by the

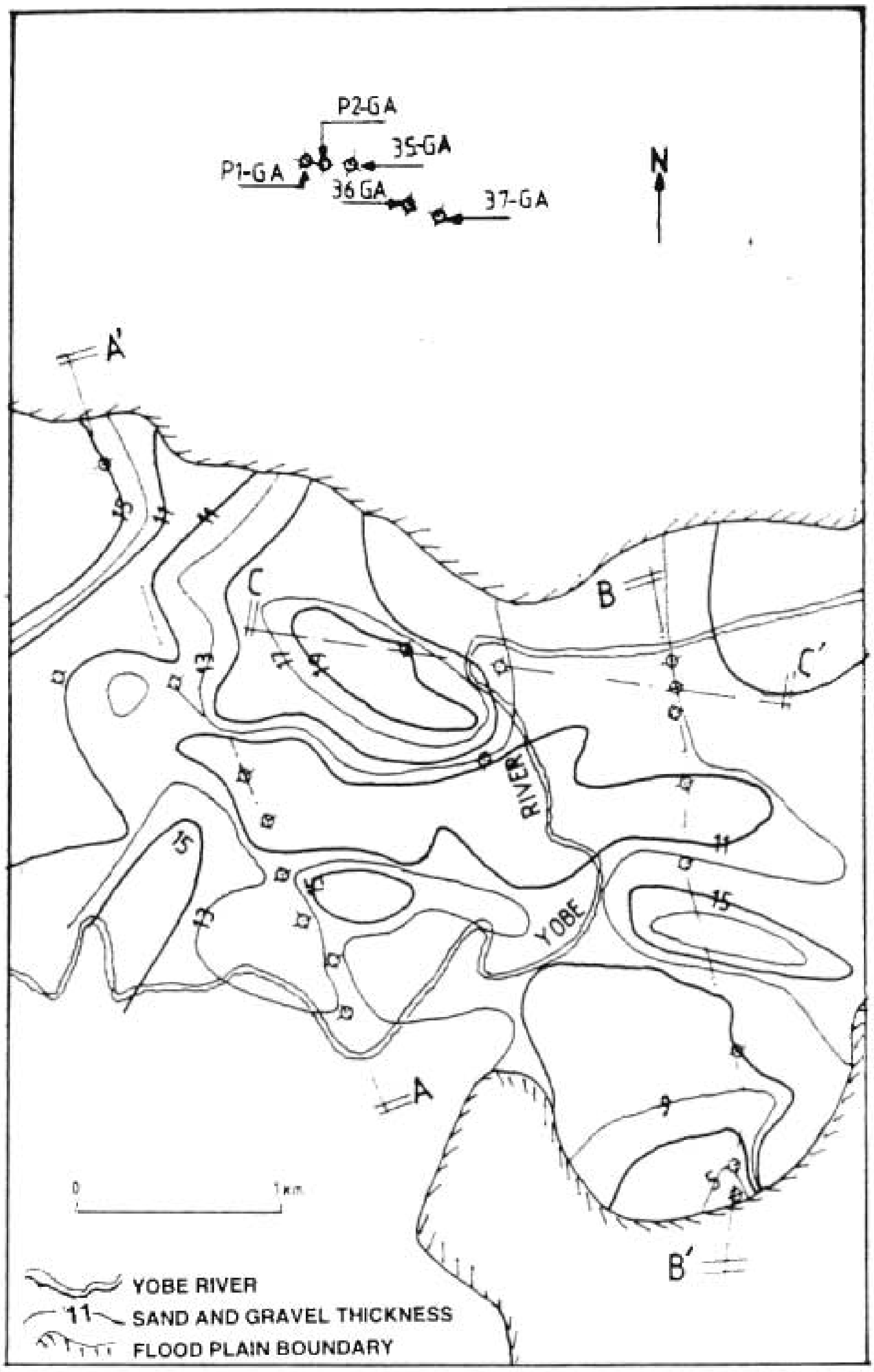


Figure 5.10. Isopach map of sand and gravel layer. Both geophysical and borehole logs data were used in the construction of the map.

thickness of the upper clay unit. For example the thickness of this unit between P3 and P5, and near the southern margin of the floodplain is reduced by the top clay unit (section B-B').

Within the central portion of the floodplain, a strip of the sand and gravel unit, 11 metres or less thick, trend in the west-east direction and can be traced over a distance of up to 3km. North of this strip, as indicated by the contours, the sand and gravel unit thickness increases significantly over a short distance and thicker deposits of sand and gravel are encountered (section C-C',figure 5.9). The sand and gravel deposits encountered at borehole K was 19.50m thick, which is the maximum thickness obtained for this unit in the study site during drilling. From this depression towards the north-west the thickness of the sand and gravel unit decreases gradually until it begin to increase again near the north-west corner of the study site. Two other depressions can also be seen south of the central sand and gravel strip. In these latter cases the sand and gravel unit is about 15.0m thick. The sand and gravel unit thickness decreases towards the southern margin of the floodplain.

An interesting observation from the map of saturated thickness is that the permeable sand and gravel unit does not pinch out along the north and south margins of the floodplain but rather maintains considerable thickness, especially near the northern margin of the floodplain. Drill logs for water supply boreholes located in the uplands at Gashua town indicate an overbudden of 3m to 4m of clay bound material underlain by relatively thick (18 m to 26m) deposits of fine to coarse-grained sand (figure 5.11). These boreholes are located at about 1750m away from the northern edge of the floodplain (figures 5.8 and 5.10). Based on this determination, it appears that the middle Yobe valley aquifer, at least within the study site, is in lateral hydraulic continuity with permeable units underlying the uplands, and therefore, is not an isolated system. Presumably the lateral continuity of both the upper clay unit and the sand and gravel below would suggest a formerly more extensive floodplain, or a floodplain which has migrated to some extent. However, further detailed field work is necessary before the relationship can be determined between the alluvial aquifer and the complex Chad Formation system underlying the uplands.

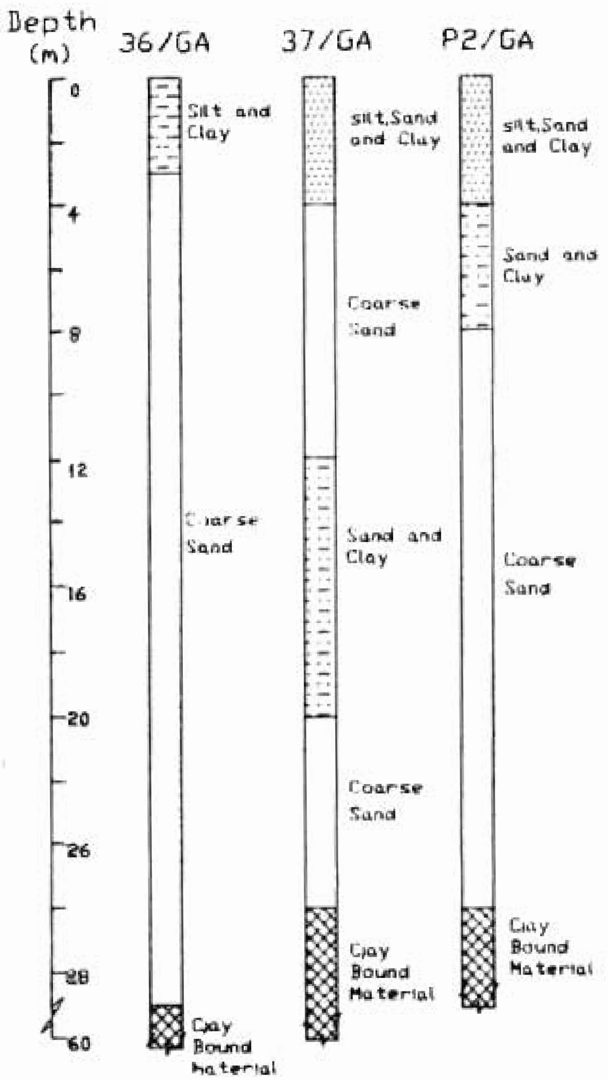


Figure 5.11. Borehole logs in the upland (Gashua Town)

The Underlying Clay Layer

This layer is continuous and forms the base of the overlying sediments. The surface of this layer contains both depressions and ridges, and these features at some locations have influenced the distribution of the overlying sediments.

Sediments' Depositional Model

The types and dispositions of sediments encountered in the present study are not unexpected from a fluvial environment and reflect the inherent complexity of fluvial deposits caused by the intricate interactions of numerous variables including stream gradient, bed load/suspended load ratios, floodplain dimensions, channel mobility, and climate (Schumm,1977; Leeder,1978; Bridge and Leeder, 1979). Galloway and Hobday (1983) have given indications of the types and dispositions of fluvial sediments that might be encountered under different channel conditions. These range from bedload deposition in straight, high energy channels to load deposition as point bars and chutes in meandering and anastomosing, low energy environments.

The Yobe River system changes from a relatively straight channel in its upper reaches to slightly sinuous in the middle section, to highly sinuous in its lower sections. Hazell et al (1988) have shown the similarities between Galloway and Hobday's model and the types and dispositions of sediments encountered along the middle section of the Jama'are River. By combining available borehole logs and data from resistivity and electromagnetic surveys, they produced a map (figure 5.12) showing the pattern of alluvial deposition along the middle section of the Jama'are River. The middle section shows significant evidence of change from meandering channels to a modern low sinuosity channel stage. In general terms, the alluvial depositional pattern obtained fall into one of the three categories of Galloway and Hobdays' model. The categories are:

1) low sinuosity channels: Low sinuosity channels comprise modern river channels and older (or overspill) straight channels of coarse sand and gravel frequently overlying clays. Bedload sands and gravels deposited as longitudinal and transverse bars and lag point, in turn form a continuous sand-filled channel with high width to thickness ratio. A multi-stage development of the river is also possible resulting in upward fining and complex bedding associated with repeated erosional and depositional phases.

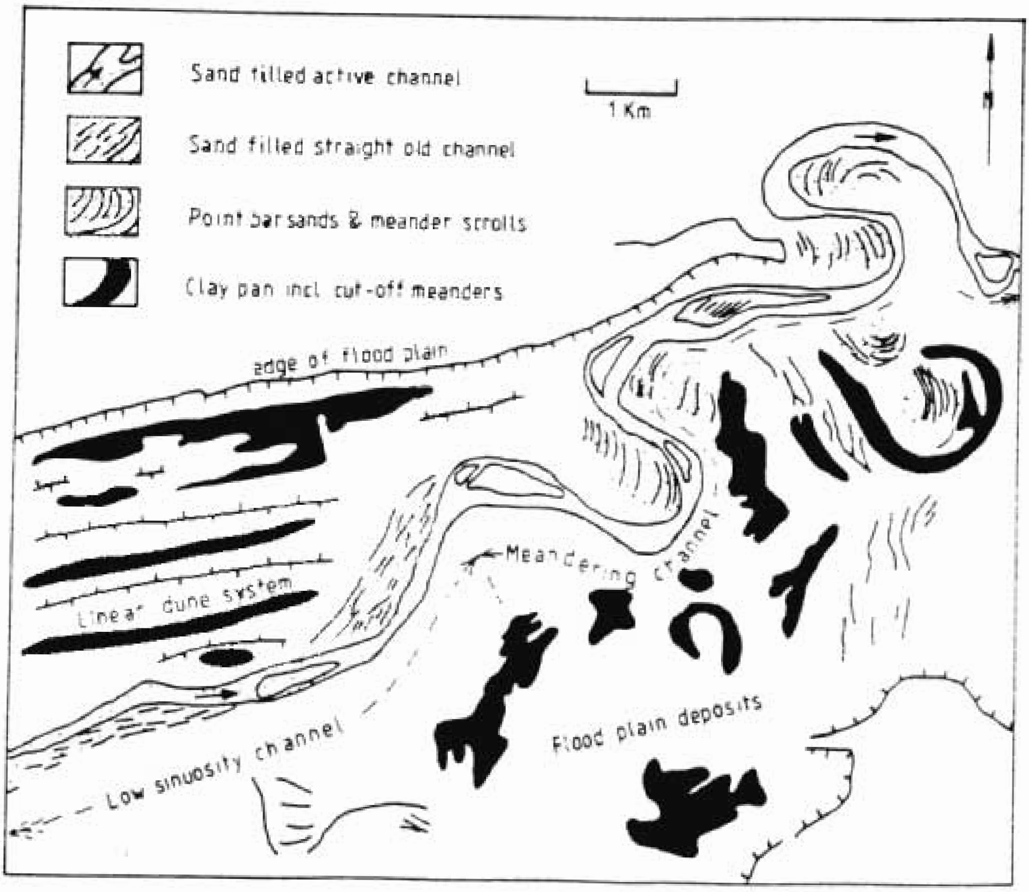


Figure 5.12. Alluvial depositional environments near Zubuki, Jama'are River (After Hazell et al., 1988)

2) meandering channels: Galloway and Hobday's possible depositional model for a river of medium to high sinuosity has two main components: coarse channel units formed by sand and gravel carried by tractive transport and deposited by lateral accretion on point bars on the inside of each meander, and the migration of the meanders in the past has resulted in a fairly complete spread of sand across the valley; and floodbasin units formed by vertical accretion of sediment from suspension in flood waters. The channel and floodbasin units alternate, together comprising fining-upward cycles. Channel units commonly start with an erosional base covered by lag

gravel, passing up to sand, which decreases in grain size upwards. Surfaces of point bars frequently have mud-filled meander scrolls which can be recognised on air photographs (figure 5.12). Chutes, chute bars and cut-offs are common features in the meander belt as the river migrates across the floodplain. Identification of coarse sand deposits means essentially the recognition of point bars and channel lag deposits in abandoned channels and chutes. Depending on the thickness of pan or floodplain deposits overlying these features, they may or may not be recognised on airphotographs.

3) overbank (pan) deposits and relic meanders: Overbank (pan) deposits and relic meanders occur extensively in the middle Jama'are valley. Formed by overbank flooding and deposition of silty clays both on alluvial flats of the floodplain and in the abandoned meander channels, pan eventually forms a masking layer over former mud plugs and sand bars.

The above model may also be used to describe the types and dispositions of the sediments encountered in the present study. The variation in the compostion and the upward fining of the sediments observed within the study site reflect the changing river conditions under which the sediments were formed, while the appearance of troughs and ridges on the surface of the underlying clay layer indicates the locations of the channels abandoned by the Yobe River (figure 5.9). It is plausible, for example, that the depression encountered around borehole K (figure 5.10 and section C-C') was formed as a result of the Yobe River cutting-off its meander either by flowing through a swale on a point bar (chute cut-off) or by breaching the neck of its meander (neck cut-off), or by avulsion. The degree of change in the river course increases in the order described above, as does the length of the channel abandoned. The abandoned channel was gradually filled-up by vertical accretion deposits during subsequent floods, forming a silt and clay layer of 1.5m thickness which is presently resistant to subsequent lateral erosion, thus exerting a degree of control on the location of the present river channel. The present Yobe River turns sharply northeastwards adjacent to this depression and flows towards the east in a straight channel. A similar explanation may also be given for the other depressions observed in the study site. For example the trough under boreholes A, B and C (section A-A') was formed as a result of the present river shifting to a new position south of this depression. The abandoned channel was filled up with 5.0m of silt and clay deposits which appear to take the shape of the abandoned channel. This is a classical example of a meander loop plugged and filled with fine materials. The mainly fine grained deposits formed in a meander loop are curvilinear in plan and about as broad and thick as the original channel was wide and deep (Allen,1970).

Comments and Conclusions

Geoelectric depth soundings, when used in conjunction with drill log data, can partially replace the more expensive method of drilling to obtain groundwater information. The general disposition of sediments given by the drilling reports agrees closely with those determined using geoelectric data. The drilling reports and the geoelectric surveys both show that the Yobe alluvium contains a sand and gravel layer sandwiched between two clay bound units.

The conjunctive use of resistivity and drill log data has provided valuable informations on the hydrogeologic framework of the alluvial deposits that could not be easily obtained by either method alone. For example the physical description of cuttings and "feel" during drilling give direct evidence for sediment type, but this method could not be used to determine aquifer thickness at every desired location especially at a close spacing because of technical difficulty and cost of drilling boreholes. Geoelectric sounding on the other hand could be conducted at a close spacing, but the results obtained from this method could only indicate the sediment type that may be encountered and the aquifer thickness obtained could be inaccurate unless calibrated with drill log data.

The success of the resistivity survey in delineating the major units of the Yobe alluvium could be predicted because expected field conditions satisfy the criteria for a successful resistivity survey. Despite a relatively coarse spacing for the survey, the thicknesses estimated for the different units of the alluvium are strongly correlated to those inferred from drilling. The reasons for this agreement are:

1) geologic conditions are favourable. Contacts are nearly horizontal and resistivity contrast exist between the individual units of the alluvium; and

2) borehole logs are available to calibrate the geoelectric models.

The potential usefulness of a resistivity survey to a groundwater study in the basin can be determined before the survey is undertaken. A basic consideration is whether the structures resolvable by geoelectric survey are significant in relation to the hydrologic regime of the study area. An assessment of geologic conditions and expected resistivity contrast can prevent spending time and money on a survey which will yield unsatisfactory results. If care is taken in determining the resistivities for the individual units of the sediments, reasonable quantitative interpretations can be made within limits imposed by the VES station spacing, using simple geologic models. The geoelectric survey together with available hydrogeological data facilitated the construction of the conceptual model of the Yobe basin sediments at an early stage of the present study. This enabled the subsequent drilling programme to be planned and directed in a more efficient and constructive way.

Note, however, difficulties can be encountered in using geoelectric technique in aquifer delineation. The main problem still standing in this approach is related to the inherent limitations of the electrical methods, i.e, the uncertainty in geoelectric parameters determination for thin units of the geological profile due to electrical equivalence (Koefoed,1979), and the difficulty in distinguishing between clay content, porosity, and salinity effects on the sediments' resistivity values.

Because of the rapid increase in the development of the alluvial groundwater of the Yobe basin for small scale irrigation in the region, there must be a commensurate effort to obtain detailed geologic information. The reliance on qualitative grain size and resistivity data for lithofacies definition used in the present study limited the overall values of lithofacies as predictive tool. Future lithofacies analyses (both at the study site and elsewhere) will be enchanced by: (1) incorporation of geologic parameters more closely related to hydraulic properties (e.g quantified grain size distributions) in addition to parameters relevant to geologic history which can be used to determine depositional environments, and (2) by integration of auxilliary data (e.g water levels) into the interpretation of lithofacies. The use of water level has been considered for the study site in Chapter 6.

While our understanding of the nature and disposition of the Yobe valley alluvium has increased from the combined use of geoelectric and well data, there is still an urgent need to design an effective and efficient way of combining the two approaches inorder to determine the extent and stratigraphy of the alluvium. The next step would be to find the optimal combination of the more accurate but more expensive well data and the less accurate but also less expensive geoelectric data by minimising two interrelated objectives, the cost and the observation effort. The "best" observation network is expected to have the least cost and provide the most accurate information on the spatial variation of aquifer thickness among all candidate network. In the ideal case, both the cost and the observation effort would approach zero, but obviously such a network does not exist. There are numerous techniques for selecting the network alternative closest to the ideal case. Borgardi et al.(1985), for example, used composite programming, a distance-based multi-criteria decision-making technique to find the best compromise observational network considering a set of conflicting objectives of network design. For the purpose of delineating the extent (both laterally and vertically) of the Yobe basin alluvium regionally, a similar approach could be adapted.

Conclusions may be summarised as:

1) The study has shown that at any particular location many, but not necessarily all, of the strata in the sequence may be present and that both the vertical and horizontal extents of the various strata are variable. Failure to recognise the nature and variability of the stratigraphical sequence within the Yobe alluvium can greatly increase the cost of borehole construction and the rate of well failures;

2) the results and experience of the present study demonstrated the advantages of the conjunctive use of drill log and resistivity data in investigating the Yobe River alluvium. The capability of the geoelectric survey to resolve and provide detailed picture of the basin geology is something that, given time constraints, would not have been possible with drilling alone;

3) resistivity survey generally gives layer thicknesses which are in close agreement with those determined by using the more expensive and time consuming test hole drilling. Errors in thicknesses calculated from the resistivity data may occur at locations where sand and gravel layer is intercalated with clay bands or where there is insufficient contrast between two adjacent units. Geoelectric survey may not be unique or diagnostic by itself, logs from drill holes should be used to check and calibrate geoelectric interpretations, because they provide the final, most detailed description of aquifer location and character; and

4) the procedure and data presented here are general and can be a basis for data collection and/or observation network design to determine the optimal combination of well data and geophysical measurements.

5.5 Pumping Test Result.

To obtain field values for the aquifer hydraulic parameters , a pumping test was conducted in the detailed study site using a well screened in the sand and gravel unit. The type and the construction of the pumped well, the pumping test procedure adopted for the present study, and the time interval over which drawdown/recovery data measurements were made have been described in chapter four. The collated drawdown/recovery data are given in Appendix C, and the analyses and interpretaions that follow are based on these data.

The drawdown/recovery data so obtained can be used to estimate the aquifer parameters, i.e transmissivity and storativity by employing one or more of the available methods for data analysis such as Theis (1935)'s method, and Cooper and Jacob(1946)'s technique. Detail descriptions of pumping test procedures and different methods of data analysis are given by Kruseman and de Ridder (1990). The drawdown data obtained from the pumping test were analysed according to Jacob's straight line method (Cooper and Jacob,1946). In this method time-drawdown data are plotted on semi-logarithmic paper, with time on the log scale and drawdown on the arithmetic scale. The Jacob method is based on numerous assumptions, such as the aquifer is horizontal, isotropic, homogeneous, uniform in thickness, infinite in extent, and confined by impermeable beds. If the aquifer characteristics are in accordance with these basic assumptions, then the time-drawdown data will fall on a straight line when plotted on the semi-logarithmic paper. Deviations from a straight line plot are often used to delineate boundary conditions or aquifer recharge/dewatering. In the present study drilling results indicate that the aquifer is not homogeneous and of variable thickness. Inspite of these complications, however, the Jacob method of aquifer analysis was found suitable for obtaining useful estimates of the transmissivity and storage coefficient of the aquifer. By restricting the analysis to the suitable segments of the drawdown graph the effects of deviations from the assumptions stated above were minimised.

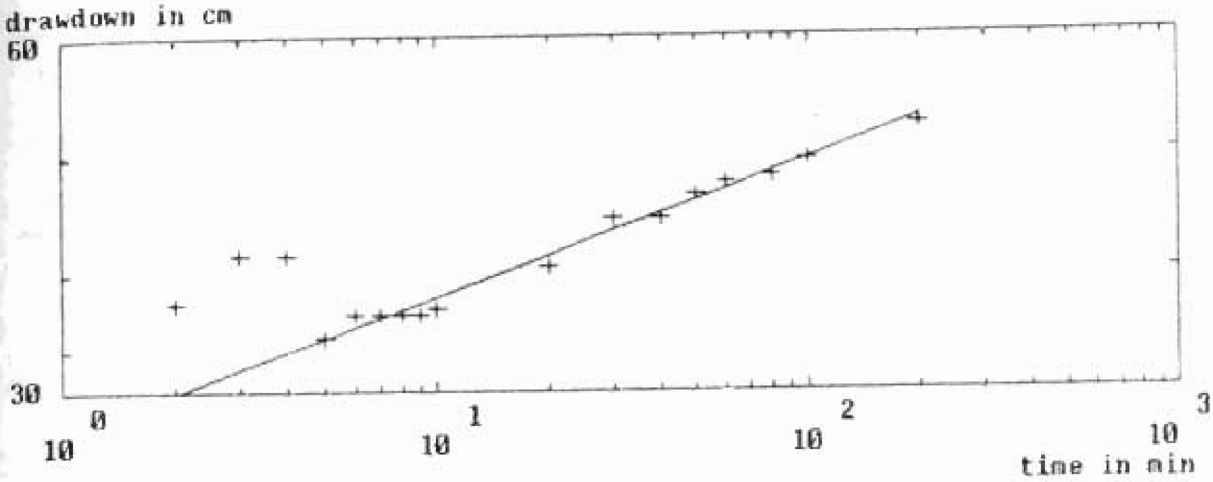
The late drawdown data from the present test generally fell on a straight line when plotted against time making application of Jacob's method reasonable (figure 5.13). Figure 5.13a shows drawdown graph with straight line graph segment at time of about 10 minutes after the beginning of pumping. Linear drawdown segment such as this was used in the Jacob method. The formula for non-equilibrum conditions as given by Cooper and Jacob (1946), data of Figure 5.13a , and measured constant pumping rate of 2.0l/s were used to calculate the transmissivity (T) and storage coefficient (S) of the alluvial aquifer.

Further, the aquifer hydraulic properties were also determined using water level recovery data. Water level measurements made during the recovery period provide a distinct set of information for the aquifer, thus providing a means of checking the results that were obtained from the analysis of time-drawdown data. Water level recovery data are often more accurate than time-drawdown data, since the recovery period is not affected by fluctuations in pumping rate. There are two common methods that are used to analyse water level recovery data. In the first method, calculated recovery versus time after pumping stopped is plotted on a semi-logarithmic paper. In the second method, residual drawdown versus t/t' is plotted on a semi-logarithmic paper, where t is the time since pumping started and t' is the time since pumping stopped. The first approach was used in this study (figure 5.13b).

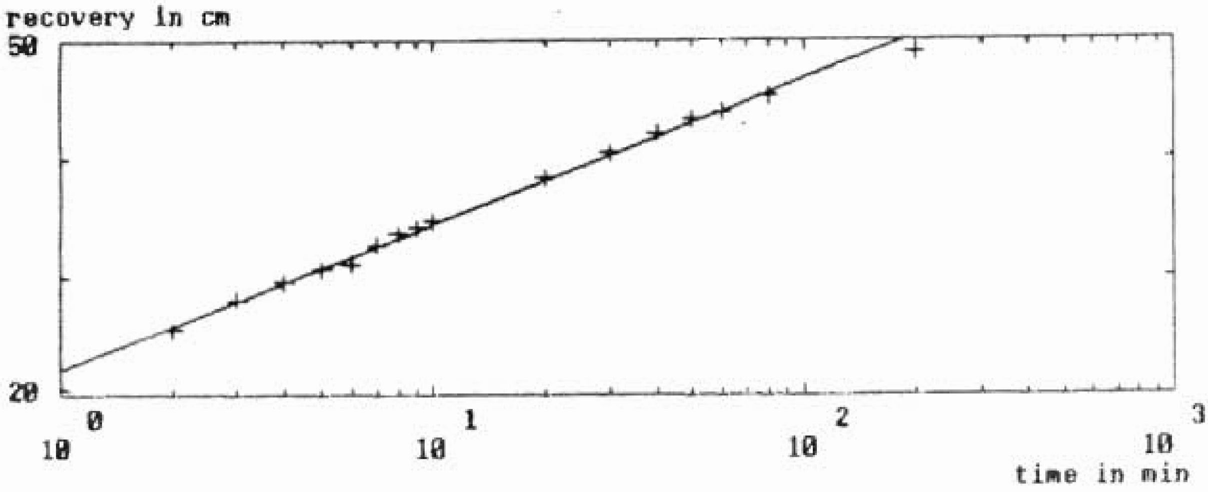
All analyses were done using a microcomputer program, SATEM, for pump test analysis (Boonstra, 1989).

Table 5.6 contains the results of the Jacob analyses of short duration constant rate pumping test and recovery data. The aquifer storage coefficient determined ranges between 1.3×10^{-4} and 2.2×10^{-4}. The storage coefficient is thus seen to fall within the typical range of values for confined aquifers. The aquifer transmissivity obtained using the two data sets ranges from 149m^2/day to 217m^2/day. The results listed in table 5.6 also indicate that the pumping and recovery data produced compatible values of transmissivity for the respective pumped and observation well. The transmissivities obtained from the observation well are 217m^2/day and 202m^2/day during pumping and recovery respectively. However, transmissivity (pumped well) is slightly lower than transmissivity (observation well) for both pumping and recovery periods. The

(a)



(b)



(Diameter of pumped well = 100mm; pumping rate = 142m^3/day; observation well distance = 4.30m)

Figure 5.13. Jacob's analysis of pumping test data in observation well. (a) Time-drawdown plot; (b) Time-recovery plot

Table 5.6. Short-term pumping test results obtained using Jacob's method

	Observation well		Pumped well	
	Pumping	**Recovery**	**Pumping**	**Recovery**
Transmissivity, $T(m^2/d)$	217	202	149	182
Storativity, S	1.3×10^{-4}	2.2×10^{-4}	--	--

transmissivity derived from the pumped well during pumping might have been affected by early drawdown variability (possibly due to well bore storage), short term effects in the most perturbed zone of the aquifer (more turbulence and non-darcy flow), and/or by late time pumping variation as the water level in the pumped well neared the maximum suction depth. In a sense, this confirms the common opinion (Kruseman and de Ridder, 1990) that it is highly preferable to evaluate the aquifer parameters from the drawdown measured in observation piezometers, because they are not influenced by anomalous properties in the vicinity of the pumping well. Furthermore, it was recently demonstrated by Butler(1990) that "the further an observation well is from the pumping well, the less the drawdown is impacted by the properties of material in the immediate vicinity of the observation well".

The hydraulic parameters obtained in the study site appear reasonable for a confined sand and gravel aquifer, but may not necessarily be representative of the entire Yobe valley alluvial aquifer. This is principally because the test data are of short duration and site specific. It should be noted that a pumping test of about 3-hours duration merely determines the response of that portion of the aquifer influenced by the test for the period of pumping. The test also determines the response of the aquifer for the particular time of the year during which the test was conducted. This is particularly important in a region where aquifer recharge is seasonal, as is the case in the Yobe basin. Therefore, a test conducted in the wet season may give different results than those of a test run in the dry season. This is particularly true in local groundwater flow systems which have limited groundwater storage and a fairly large change in groundwater storage in a given year with respect to total storage even under natural conditions (i.e, no pumping).

The ideal method of determining the sustained yield of the Yobe valley aquifer would be to run a long term pumping test of several months' duration so that the aquifer response for this duration could be determined accurately. However, in most situations, long-term pumping tests are not feasible because of technical and economic difficulties and short-term test are often adapted, the results of which are extrapolated into the future to determine sustainable yield for a well.

Table 5.7. Aquifer hydraulic parameters deduced from regional studies

Statistical parameter	Jama'are valley (Water Surveys, 1986)		Hadejia Valley (Diyam, 1987)	
	Transmissivity, $T(m^2/day)$	Storage coefficient, S	Transmissivity, $T(m^2/day)$	Storage coefficient, S
No. of data points	6	6	11	11
Mean	926	0.15	612	6.7×10^{-3}
Minimum	300	0.05	116	2.2×10^{-6}
Maximum	2000	0.28	2122	0.17

Table 5.7 gives a summary of the aquifer parameters derived from pumping tests conducted along the valleys of the Jama'are (Water Surveys,1986) and the Hadejia (Diyam,1987) Rivers respectively. The Jama'are valley aquifer transmissivity ranges from $300m^2/day$ to $2000m^2/day$ and the mean transmissivity is $926m^2/day$. The aquifer storage coefficient falls between 0.05 and 0.28, and has a mean of 0.15. Thus the mean storativity obtained corresponds to an unconfined aquifer. Similarly a wide transmissivity range can also be seen for the aquifer of the Hadejia valley. The maximum transmissivity ($2122m^2/day$) for the Hadejia aquifer is equivalent to the

maximum transmissivity ($2000m^2$/day) for the Jama'are aquifer. However, in the Hadejia valley the lower end of the transmissivity scale is $116m^2$/day, which is lower than the minimum transmissivity ($300m^2$/day) determined for the Jama'are aquifer. The storativity for the Hadejia aquifer also occurs over a wide range; storativity ranges from $2.2x10^{-6}$ to 0.17, and the mean storativity is $6.7x10^{-3}$. Note, however, that the maximum storativity of 0.17 was only encountered at one location (Gabarun town) and aquifer storativity at most locations are lower than the mean storativity ($6.7x10^{-3}$). The storage coefficients determined for the Hadejia aquifer are typically in the confined aquifer storativity range (Diyam,1987).

Aquifer transmissivity determined in the present study is thus seen to correspond to the lower transmissivity range obtained in the upper section of the Yobe basin (Hadejia and Jama'are valleys). However, it is appreciated that the aquifer parameters may vary even within the study site if several pumping tests were carried out. It would be appropriate in the future to conduct several pumping tests downstream of the present study site to determine aquifer parameters and to see if significant variations/reduction in the magnitude of aquifer parameters exists while moving downstream. A statistical analysis, such as significance test, will assist greatly in carrying out this task.

CHAPTER SIX

HYDRAULICS OF THE YOBE RIVER-AQUIFER SYSTEM

6.1 Introduction

The geologic framework of the Yobe basin alluvium of the present site has been presented in chapter five. Chapter six describes the hydraulic characteristics of the Yobe River-aquifer system.

In section 6.2 of this chapter the temporal and spatial variations of aquifer water levels recorded in 20 piezometers over a one year study period are presented to describe groundwater hydrogeologic conditions under both low and high river stage conditions. The location and type of piezometer used, as well as the installation procedure adopted, and the time interval over which water level measurements were made have been described in chapter four. The collated water level data are given in appendix B and the analyses and interpretations given in this section are based on these data.

The present work was designed primarily as a pilot study to assess the nature of the river-aquifer interaction. Geologic, geophysical and hydrological data were used in conjunction for this purpose. However, a major drawback to this approach is the likely high cost of obtaining additional information regionally if collection of the new data has not been guided by some conceptual model of the system based on the available data. Such a model would serve as the framework on which to hang details, and pertinent data can be collected without being blinded by a host of trivialities and an extra abundance of extraneous data. The third section (section 6.3) of this chapter presents a conceptual model of the system based on the results and discussions presented both in the present and the previous chapter.

6.2 Hydraulic Head Measurements.

Hydraulic Head Distribution.

To determine the spatial configuration of the aquifer water level at given times, the alluvial groundwater flow system within the study site was described with the aid of potentiometric maps and hydrologic cross sections. These maps and sections were prepared for the lowest water level in June 1992 (25/6/92), an intermediate aquifer condition in August 1992 (22/8/92, seven weeks following the commencement of the 1992 river flow), the aquifer water level coinciding with the Yobe River near its peak stage in September 1992 (15/9/92), water levels on 13/11/92 and 15/3/93, one and five months respectively, since river stage recession began, and for the lowest water level in June 1993 (27/6/93). These maps and cross sections are shown in figure 6.1a to f and figure 6.2a to c respectively. The locations of the data points upon which the contours are based are also indicated in these figures.

Six configurations of the aquifer water level are evident in the study site for these six water level conditions:

1) A map of the potentiometric surface with flow lines for 25/6/92, the end of the 1992 dry season is shown in figure 6.1a, and the hydrologic sections for the north and the south sides of the aquifer taken on this date is shown in figure 6.2a.

   An examination of the sections (figure 6.2a) indicates that the alluvial aquifer was either confined or unconfined depending on location, and whether or not it is full. A commonly observed phenomenon during drilling in April 1992 was that the aquifer water level rose to, and became stabilised at, about 2.5m to 3.5m below the ground surface within a week of piezometer installation. The rise of water levels to the stabilised positions indicated the pressure head in the aquifer and revealed the largely confined nature of the system. On 25/6/92 the aquifer was unconfined where the overlying sediment is thin (less than 2m), but confined where the sediment thickness exceeded 3m (figure 6.2a). A similar aquifer condition was

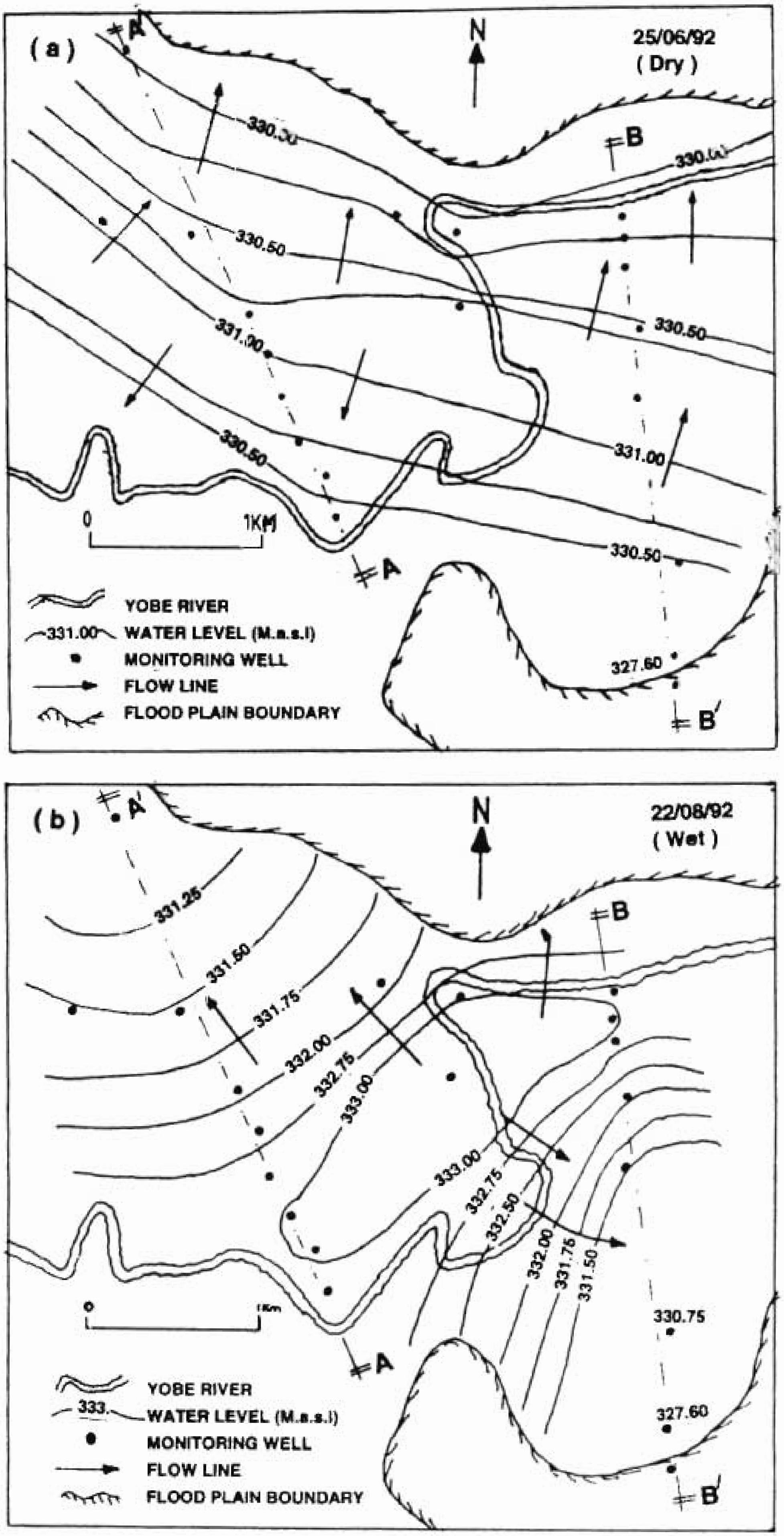
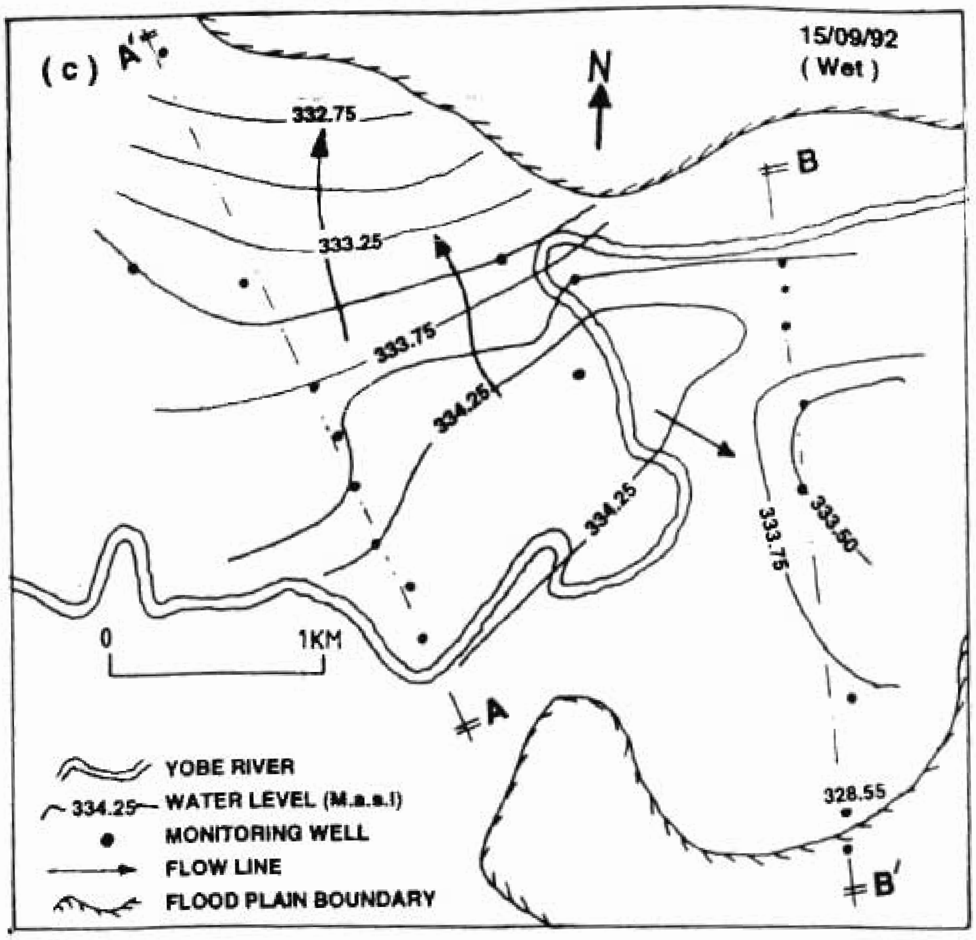


Figure 6.1. Potentiometric surface of the alluvial aquifer in the wet and dry season (1992/93). (a) configuration of aquifer water level on 25.6.92 (dry); (b) configuration of aquifer water level on 22.8.92(wet).



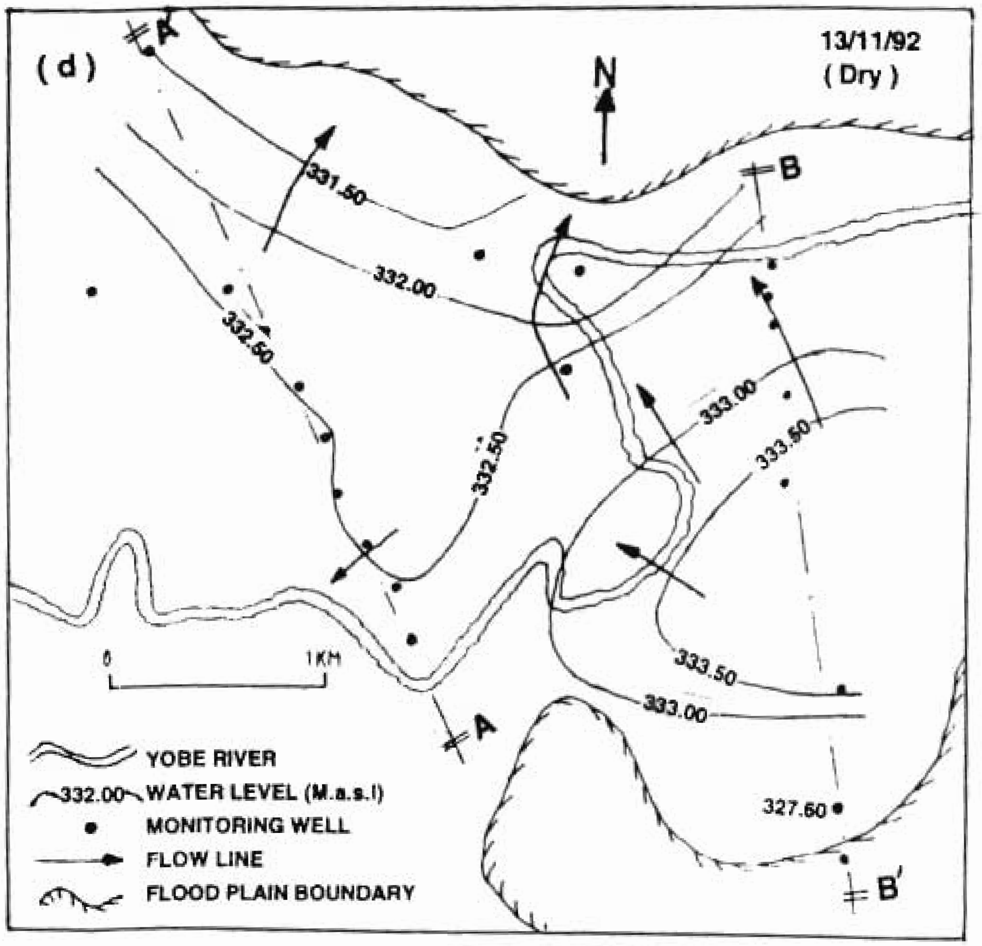


Figure 6.1. cont'd. (c) configuration of aquifer water level on 15.9.92 (wet); (d) configuration of aquifer water level on 13.11.92 (dry)

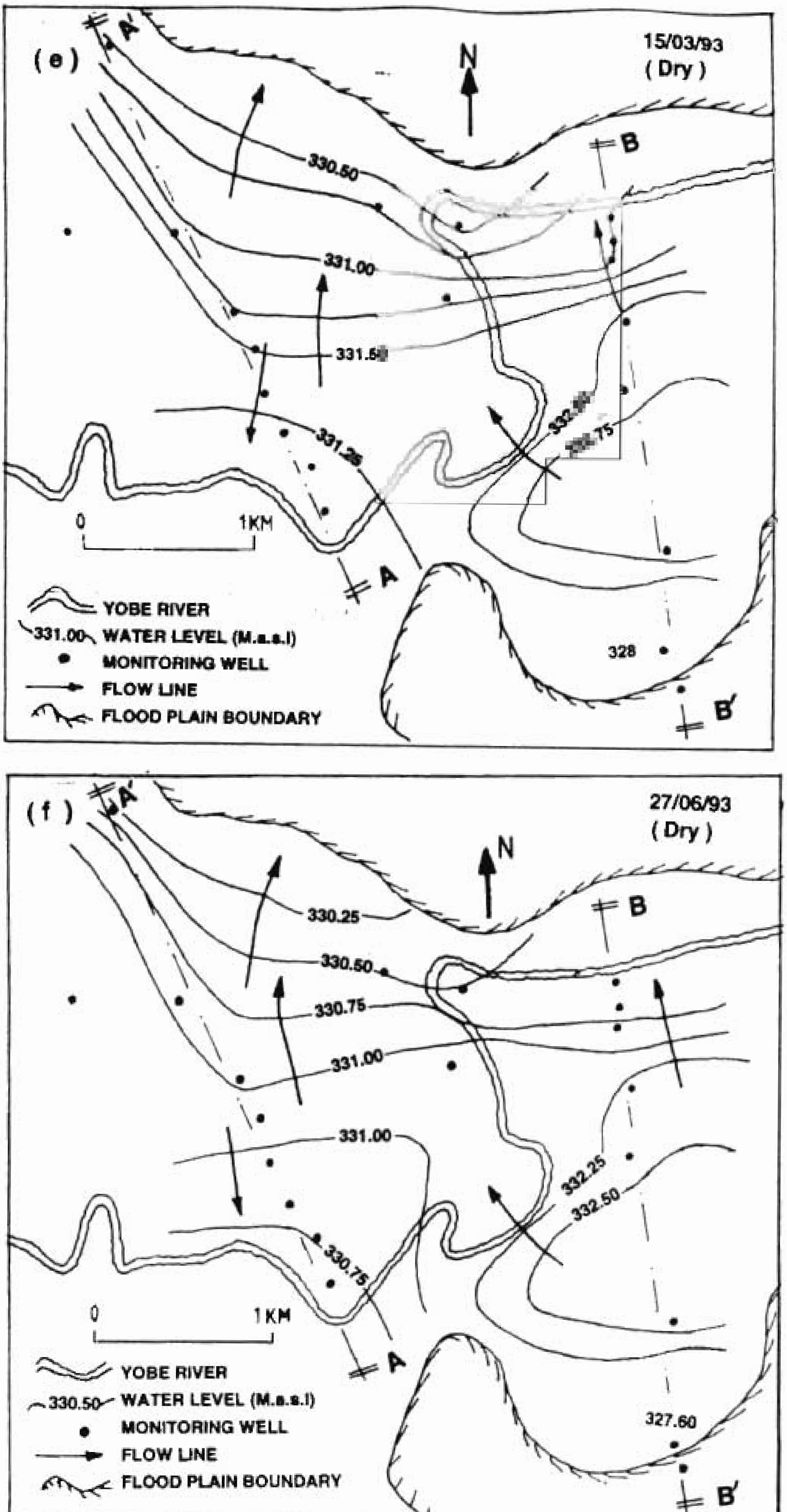


Figure 6.1. cont'd. (e) configuration of aquifer water level on 15.3.93 (dry); (d) configuration of aquifer water level on 27.6.93 (dry)

obtained on 27/6/93, except for the groundwater high in the south section (figure 6.2a).

The potentiometric map for 25/6/92 (figure 6.1a) depicts the following:

a) the water level elevation north of the river was higher (331.0 metres above sea level (masl)) at the centre of the floodplain decreasing to 330.50 masl and 330.0 masl in the vicinity of the river and the northern margin of the floodplain respectively; and

b) on the south side the water level elevations declined from 331.0 masl at the floodplain center to 330.0 masl near the river bank. Beyond the floodplain center there was a localised steepening of the hydraulic gradient towards the southern margins of the floodplain. Water level elevation decreased from 331.0 masl at the centre to 327.60 masl at the southern floodplain margin.

An assessment of the topography in the present site showed that topography has no definite control over the groundwater levels. Ground surface elevation ranged from 333.0 masl to 334.0 masl. Consequently two directions of subtle groundwater flows are discernible in figure 6.1a. On the north side some of the groundwater flows from the centre towards the river while the remainder moves north towards the adjacent upland. The flow was slow, the lateral hydraulic gradient being 4.10^{-4} and 6.10^{-4} toward the river and the northern margin respectively. Groundwater south of the river mainly flows towards the river. Groundwater flow from the floodplain centre towards the river during the dry season would be expected under natural drainage conditions.

2) The aquifer water level condition on 22/8/92 is shown in figure 6.1b. The prominent feature of this map is a NE-SW trending mound at the study site centre. Groundwater flow was directed radially from the mound and the mainly northward water flow of the dry season (25/6/92) was changed to a predominantly NW and SE trending flow. The contour lines are arcuate and more closely spaced on the east side of the mound than on the west, an indication that the groundwater flow patterns on the two sides (north and south) differ. The magnitude of flow was

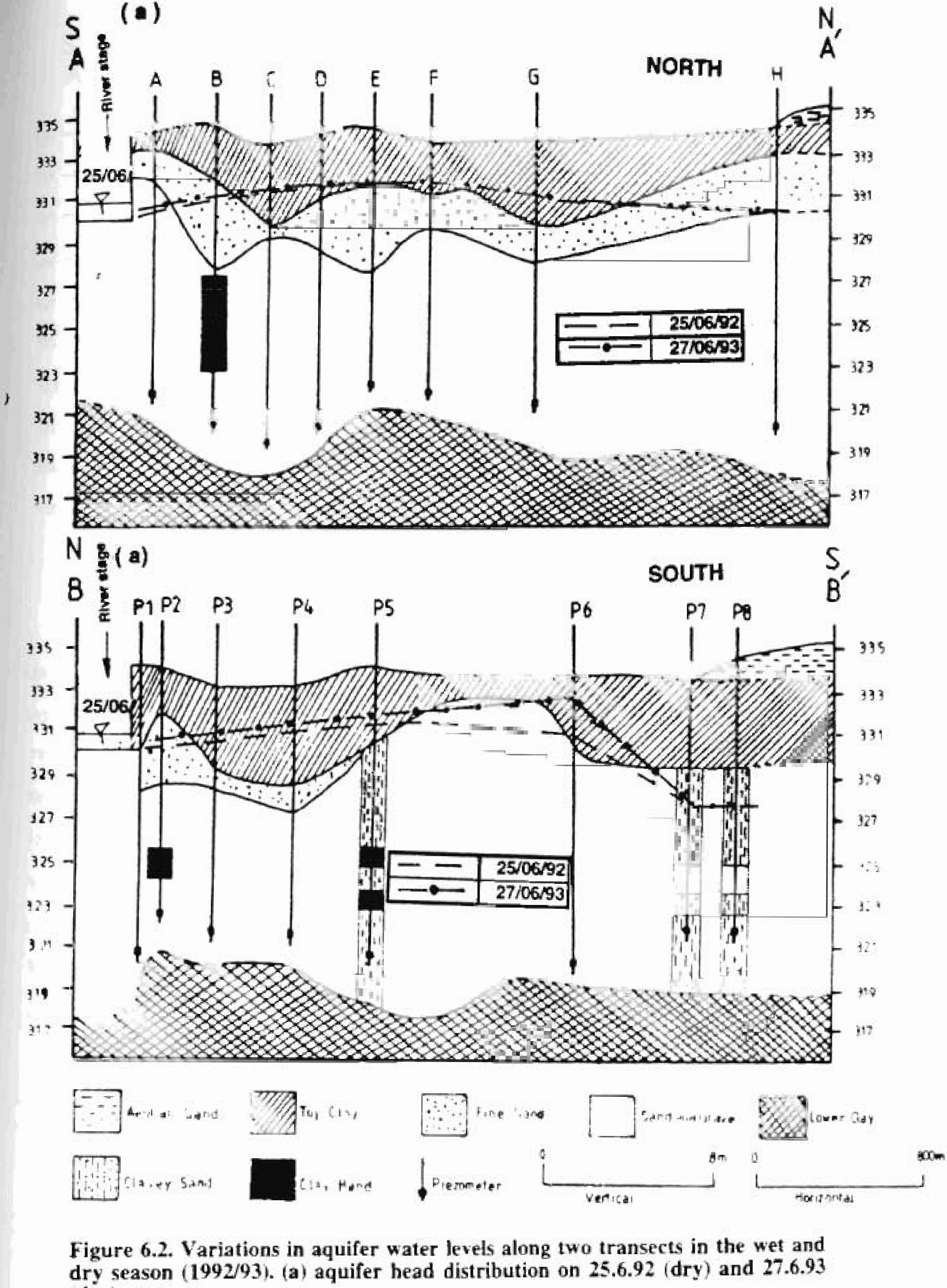


Figure 6.2. Variations in aquifer water levels along two transects in the wet and dry season (1992/93). (a) aquifer head distribution on 25.6.92 (dry) and 27.6.93 (dry)

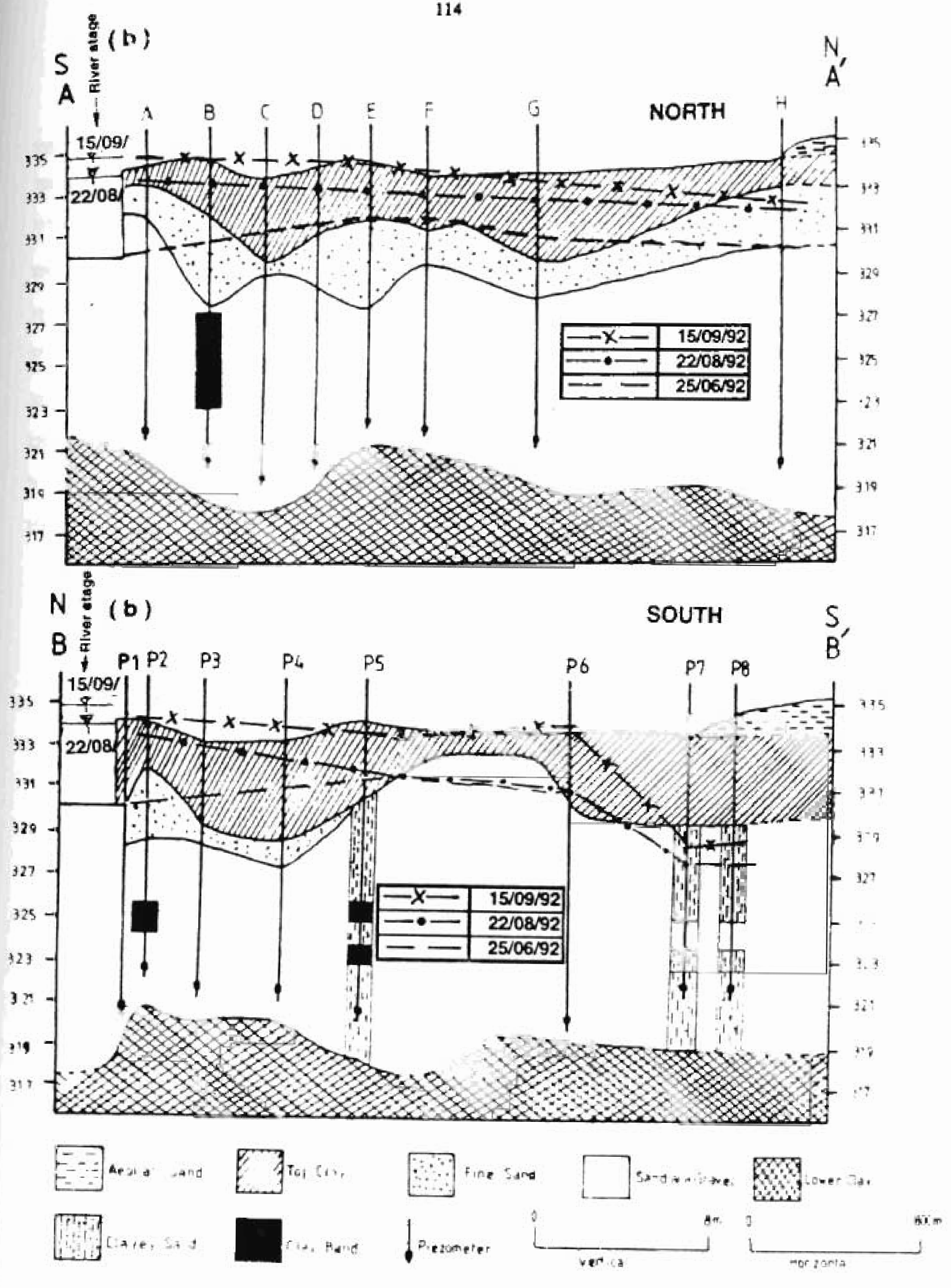


Figure 6.2. cont'd. (b) aquifer head distribution on 25.6.92 (dry), 22.8.92 (wet) and 15.9.92 (wet).

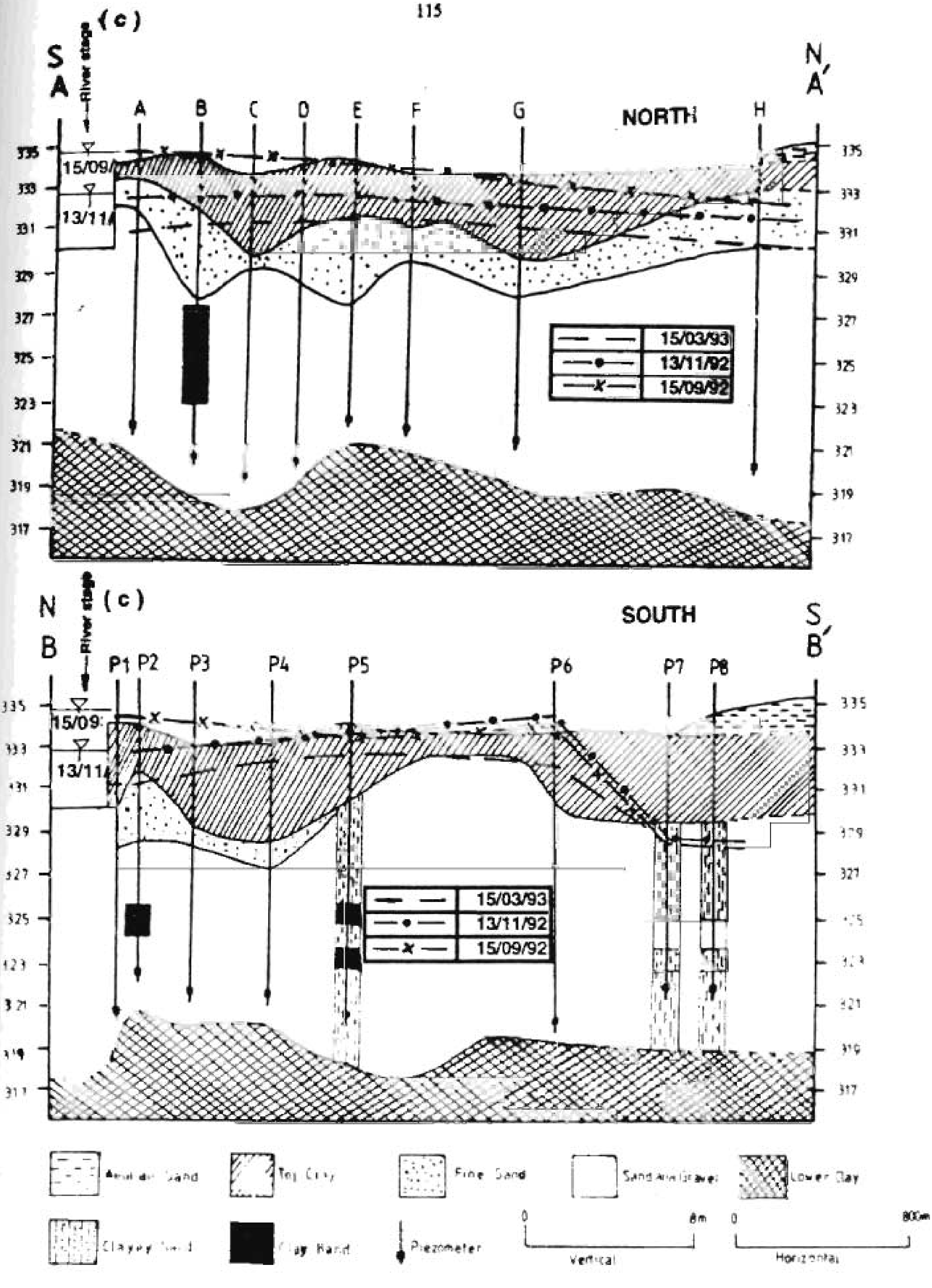


Figure 6.2. cont'd. (c) aquifer head distribution on 15.9.92 (wet), 13.11.92 (dry and 15.3.93 (dry).

larger towards the NW than SE. The groundwater mound apex is at 333 masl and declines to 331.25 masl near the north flood plain margin. The changes in water level configuration and flow direction from June to August 1992 were due only to seepage from the Yobe River channel (the river had not exceeded bankful by August). The appearance of the 331.25 masl contour, which is equivalent to the central water level high of the 1992 dry season (figure 6.1a), near the north boundary (which is about 3km from the river) in about 7-weeks of river flow indicates the rapidity with which the seeping river water was being transmitted into the aquifer. This in turn reflects the degree of the river-aquifer connection and lateral geologic continuity.

On the south side transmission was slower and the position of the 331 masl water level was shifted less (compare figures 6.1a and 6.1b). This is the result of, among others, a differing aquifer transmissivity. The cause of transmissivity variation and its effect on groundwater flow is discussed in a later subsection.

The influence of the lateral seepage on the alluvial aquifer hydraulic head distribution is shown in cross section in figure 6.2b. An examination of figure 6.2b reveals that:

a) All of the piezometers located north (section A-A') of the river, and some (four out of the eight piezometers) on the south side (section B-B'), experienced between 1.0m and 3.0m increase in water level;
b) in the section north of the river the rise in aquifer water level caused by the river flow was greater and all of the aquifer, except for a small portion near the north flood plain margin, was brought under a confined condition; and
c) in the section south of the river the aquifer response to the river flow was slower and the conversion of the aquifer condition, from unconfined to confined, was limited to areas within 1000m of the river channel. Piezometers located at distances greater than 1000m from the river (P5 and beyond) experienced

negligible increase in water levels and practically remained near their dry season (25/6/92) positions.

3) Measured water level on 15/9/92, a date on which the Yobe River stage is near its peak and the flood plain surface was mostly inundated, is displayed in figure 6.1c. A mound similar to that obtained in August, but one with an apex higher by more than 1m, was obtained; the mound apex is at 334.25 masl. Generally there was a cumulative water level increase of between 1 and 3m throughout the study site since 25/6/92 (compare figure 6.1a and c). The contour spacings resemble those of 22/8/92 but the flow directions are mainly toward the north and east. The increased river flow and subsequent flooding induced the water level near the south flood plain margin to increase by 1m from its dry season position (327.60 masl) and rose to 328.55 masl. On this date nearly the whole of the aquifer was brought under a confined condition (figure 6.2c).

4) The recorded water level on 13/11/92, approximately a month after recession began, is shown in figures 6.1d and 6.2c. The central mound of the wet season has been dispersed and the contours are widely spaced, an indication that the intensive flow of the wet season has been reduced. As a result water level in the north sector fell back by more than 1m from its high position on 15/9/92. A similar level decline (about 1m) was also experienced by piezometers (P1 to P4) located closer to the river in the south, but the change was relatively small in the central region during the one month recession period. Negligible decline in piezometric level was noticed beyond P4, and at P6 the level on 13/11/92 stood above that on 15/9/92 (figure 6.2c). Water level in the south was generally higher by more than 1m. Groundwater flow was mainly towards the north, and as in the earlier cases, flow towards the south was again restricted (figure 6.1d).

5) The piezometric contours shown in figure 6.1e depicts the aquifer water level condition on 15/3/93, five months into the recession. The flow was predominantly towards the north. Groundwater recession continued and there was a further decline in levels. The water level in the north declined from its position on 15/9/92 by

nearly 3m while in the central south the decline was about 1m. The water level in south centre only fell (just) below the upper clay. This aspect can be seen in the cross section shown in figure 6.2c.

6) The aquifer water level on 27/6/93, the end of the 1993 dry season, is shown in figure 6.1f. Two facets of the map are instructive:
   a) The disposition and magnitudes of the contour lines north of the river are similar to those obtained on 25/6/92. The effects from the 1992 flood have been reversed and water levels have returned to their 1992 dry season positions. The 1993 potentiometric surface is higher (compared to that of 1992) by about 0.25m near the north flood plain margin. The central north region is at 331 masl and was being discharged towards the river and the north flood plain margin, a condition identical to that obtained on 25/6/92 (compare figures 6.1a and d).
   b) The distribution of the piezometric surface on the south side differs from those attained on the north side, and those obtained for the south on 25/6/92 (compare figures 6.1a and f). A potentiometric surface of about 333 masl formed a mound in the central portion. The main flow was towards the river while flow was limited towards the south beyond the local mound.

The major observations and features of the maps and cross sections shown in figures 6.1 and 6.2 can be summarised as:

1) There was a general increase in aquifer water level following the 1992 river flow and flooding. Most part of this effect was subsequently reversed in the dry season; the decline was more rapid in the north.
2) there was a persistent groundwater gradient from the centre towards the north during both low and high water level conditions and water was being discharged from the alluvium;
3) there was a localised steepening of the hydraulic gradient towards the south flood plain margin during both low and high water level conditions, but little lateral flow

was occurring towards it. Generally water flow in this section of the aquifer was retarded compared to that in the north; and

4) a mound had developed on the south central section after the 1992 flood. The mound is a product of a localised vertical flood water infiltration which took place around this location during the 1992 flood. This aspect is discussed in a later subsection.

Yobe River-aquifer Interactions

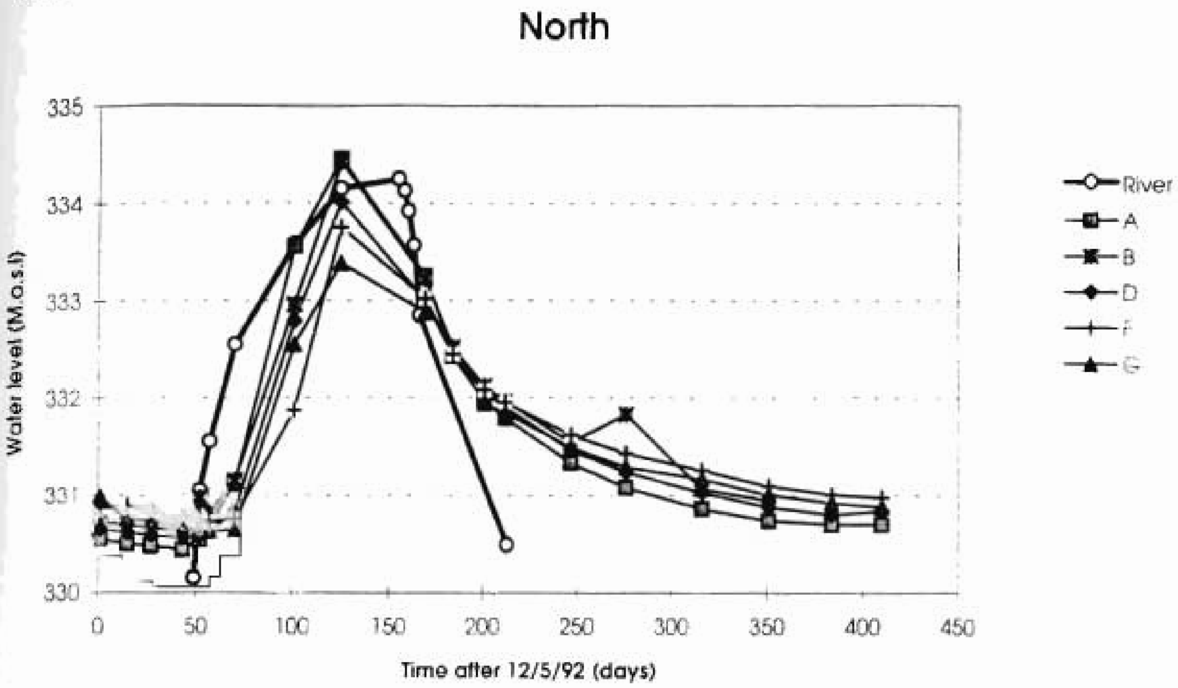
It is well known that water levels in an aquifer connected to a river can be expected to rise or fall in response to respective increases or decreases in river water levels. The ideal situation is for the aquifer concerned to have high hydraulic conductivity and direct access to the river water, resulting in the maximum aquifer response to the river stage change effects owing to the intimate relationship between flow and hydraulic conductivity.

Figure 6.3 (a and b) shows piezometric water levels prior to, during, and after the 1992 flood event. The figures shown are representative hydrographs for the 20 piezometers located on two transects, north and south of the Yobe River. In this figure water levels obtained from piezometers located at similar distances from the river such as A and P2, B and P3, are shown to make easy comparison and to establish the nature of the relationship between the Yobe River and the adjacent aquifer.

The formal recording of the piezometric water levels started on 13/5/92 approximately a month after all the piezometers were installed and time zero in figure 6.3 corresponds to May 12, 1992. The values of hydraulic head are expressed as metres above sea level at the respective piezometer location.

The influence of river stage changes on the groundwater system was studied by using the river stage data obtained from the nearest gauging station at the Gashua Bridge. The Yobe River stage hydrograph coincident with 1992 flood event is also shown in figure 6.3. The Yobe River flow conditions at the Gashua Bridge at different times of

(a)



(b)

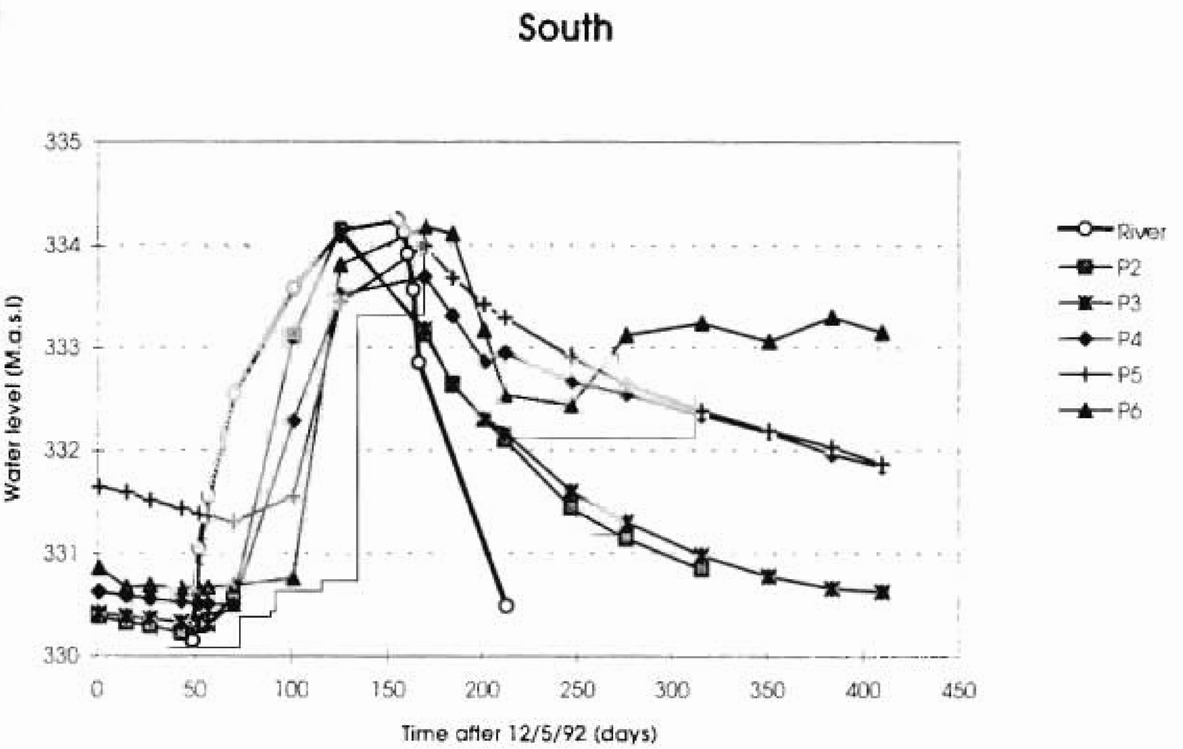
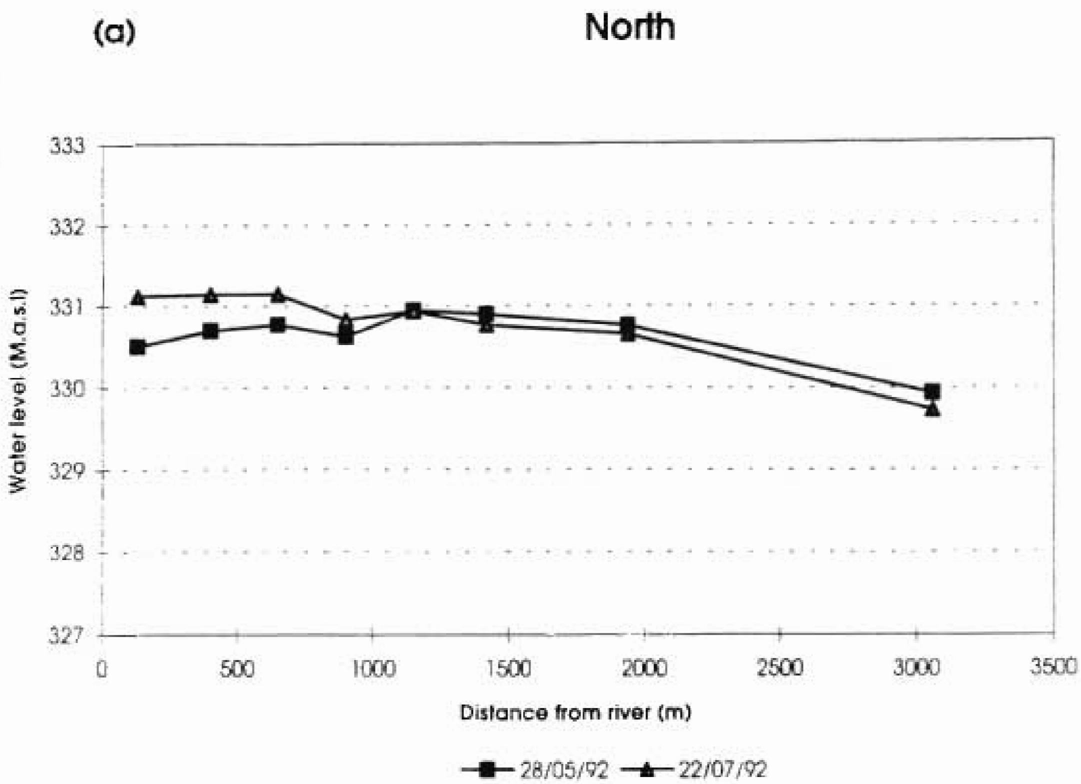


Figure 6.3. Hydrographs showing changes in aquifer water levels in piezometers compared with river stage (13.5.92 to 27.6.93). (a) five piezometers north of the river; (b) five piezometers south of the river.



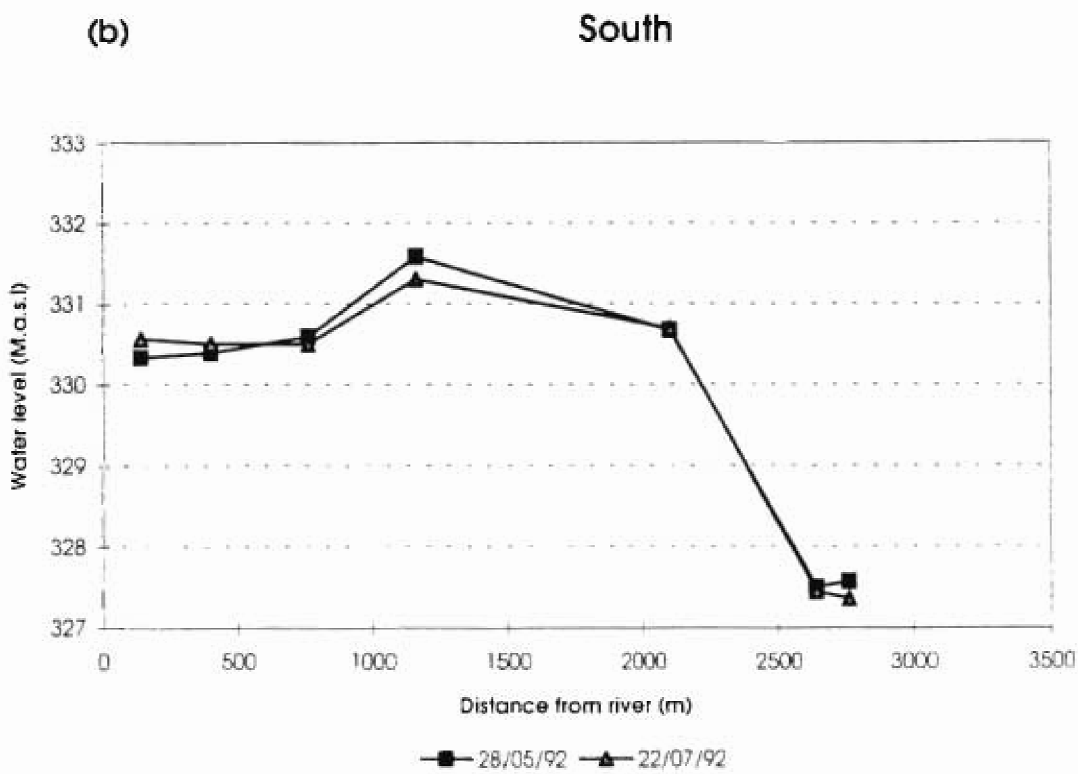


Figure 6.4. Aquifer water level response to river stage increase on 22.7.92. (a) piezometers A to H located north of the river; (b) piezometers P2 to P8 located south of the river.

the year (1992) are contained in table 6.1. The river flow condition ranges from zero flow (30/6/92) to peak sustained flow with flood plain surface inundation (15/10/92).

Flow was little more than a trickle on 30/6/92 and the river channel was empty at all locations except for the few discrete ponds such as the one under the Gashua Bridge. The river stage at the bridge was 1.30m on 30/6/92. The 1992 river flow reached Gashua on 1/7/92 and the measured river stage was 2.20m on 4/7/92, 3.70m on 22/7/92, and 4.70m on 22/8/92, the cumulative stage increase being 0.90m, 2.40m and 3.40m respectively (table 6.1). Throughout July and August 1992 the river was confined to its channel. The river stage rose to its peak of 5.40m on 15/10/92 when the river channel was full and the floodplain surface was inundated. After the flood peak the river receded rapidly and by 12/12/92 the river stage was 1.64m (table 6.1). The flow conditions and their corresponding river stages are referred to throughout this chapter.

The distribution of the 1992 flow was typical for the Yobe River, river stage rising to its peak in about three-and-a-half months of river flow and receeding rapidly. It is worth noting that the 1992 flood year (early July to mid October 1992) was particularly wet with a peak river flow of 185m^3/s, having a return period of 7-year. Consequently the subsequent discussions in this chapter, which are related to the river flow, represent an above average condition.

The river discharges were calculated using a rating equation (equation 6.1) developed for the Yobe River at the Gashua bridge by Thompson (1992).

$$Q= 2.01(H)^{2.68} \qquad (6.1)$$

Where Q= River discharge (m^3/s); and

H= river stage (m).

Equation (6.1) has R^2=93.2.

Table 6.1. Yobe River stage at the Gashua Bridge (1992).

Date	River Stage (m)	Cumulative Stage increase (m)	River flow condition
30.6.92	1.30	0	River channel is dry except for discrete pools
4.7.92*	2.20	0.90	Confined to channel four days since it started flowing
22.7.92	3.70	2.40	Confined to channel
22.8.92	4.70	3.40	Confined to channel
15.9.92	5.30	4.00	Bankful flow with flooding. Most parts of the flood plain surface have been flooded
15.10.92	5.40	4.10	Flooding was almost complete. Only few "high" grounds were dry
23.10.92	4.72	-0.68	River stage had fallen, flooding had receded and the river confined to its channel again. Floodplain mainly drained; only few depression storage remained.
12.12.92	1.64	-3.76	Stage had returned close to its dry season position

Note

(1)* River stage increase started on 1.7.92 but no stage measurement was made.

(2) River channel depth ranges between 4.0m and 5.0m with a mean value of 4.30m.

Comparison of measured groundwater levels with river stage led to the following basic observations:

1) groundwater levels are influenced by river stage changes; piezometeric levels rise or fall in response to respective increases or decreases in river stage. The piezometric level responses, except for water levels in piezometers P4 to P8 south of the river, coincide and increase with river stage increase;
2) the measured water levels in most of the "quick" responding piezometers were above the ground surface on 15/9/92 (day number 125) even before the peak river stage was recorded on 15/10/92. The peak water levels in P4 and beyond occurred on 29/10/92 (day number 169), after about 15 days into the 1992 flood recession and are therefore out of phase with peak river stage; and
3) the distance from the river has an influence on groundwater levels, the response being greater and more rapid closer to the river. Piezometer A, for example, located at about 130m from the river bank responded rapidly to the river's rise in stage, while the pulse of the seeping river water was not experienced in piezometers located at distances greater than 1000m from the river even after three weeks of river flow (figure 6.4).

The early increase in river stage in July 1992, though relatively large (the river stage increase by 22/7/92 was 2.4m out of a total stage change of 4.30m), has little impact over the groundwater system except for reversing the hydraulic gradient near the river. Following the commencement of river flow, the hydraulic head at the river bank exceeded that in the adjacent aquifer, and while the gradient adjusted to the river level increase, water movement in the aquifer immediately adjacent to the river was reversed; the effect of this was to reduce the groundwater outflow and to cause a storage in the aquifer which would otherwise have flowed to the river. This gradient reversal extended to about 1000m and 700m inland on the north and south sides of the river respectively (figure 6.4). The water level north of the river was relatively flat beneath the interior of the floodplain, but the gradient increased outward (beyond 2000m from the river) towards the northern margin of the floodplain (figure 6.4a).

The piezometric levels responded sharply to the river stage increase in August and September 1992. During this period, in almost all cases, there was a significant increase in groundwater levels. By 22/8/92 the river stage increase was 3.40m but the Yobe River was still confined to its channel. Consequently the observed increase in groundwater level from the start of river flow up to this date was due entirely to lateral channel seepage and this allowed an examination of the aquifer response to, and its characteristics under, the channel seepage stress. The north section of the aquifer was more responsive to this input than the south. This aspect was discussed earlier and represented in figures 6.1b and 6.2b.

By mid September (15/9/92) the river stage rose to 5.30m, the river channel was full and most parts of the flood plain surface were inundated (table 6.1). By this date significant increase in groundwater level was recorded and most piezometers experienced more than 3.0m water level increase. This led to the conversion of the unconfined portions of the south aquifer to a confined state and establishment of a nearly "full aquifer condition" with piezometric levels at most locations rising above the ground surface (figure 6.2c). Thus it can be said that significant groundwater level increase in the south sector of the aquifer takes place during intensive and sustained river flow conditions, while the early river stage increase mainly sets the engine for groundwater level increase working.

A plausible explanation for the differing response by the two sections of the aquifer to the earlier changes in river stages can be postulated as follows. At the early stages of the river flow most of the south section of the aquifer was unconfined with high storage coefficient, several orders of magnitude larger than that of a confined aquifer,and hence a small increase in groundwater level would require a much larger flux of water than in a confined aquifer because of aquifer storage. As a result groundwater level changes cannot take place within the same time scale as those in the confined north aquifer since the seepage waves were damped, and this has the effect of slowing down the seepage flux transmission. The seepage flux transmission in a confined aquifer connected to a river which experiences a rise in stage is so rapid that Sophocleous(1991) suggested viewing the seepage pulses as adjacent fluid molecules bumping into each other like dominoes falling in a row, or as the propagation of sound waves. The distance the molecules move to transmit the pressure is very small.

At low river stage piezometric water levels were above river water level and the existing hydraulic gradient was driving flow from the aquifer into the river. For example, between 13/5/92 and 1/7/92, a length of time covering the recorded 1992 dry season (day number 1 to 50 in figure 6.3), piezometric levels south of the river (figure 6.3b) were at the same or higher elevation than the river water level and groundwater was supplying baseflow to the river. Presumably the pool of water under the Gashua Bridge (observed during field work) was derived from this groundwater outflow. Groundwater levels north of the river also show a similar trend (figure 6.3a), except that the river channel along this length of the river perpendicular to this transect was dry, even though groundwater outflow was seen to be taking place. Note that the river stage hydrograph shown in figure 6.3 is that obtained at the Gashua Bridge and the river channel was dry at almost all locations before the onset of the 1992 river flow. This observation may suggest the existence of a groundwater flow component parallel to the river channel which controls the movement of the water discharged from the aquifer. This flow component was likely being controlled by the slope along the channel bed and a point of discharge to the land surface within the channel somewhere downstream. It is possible that the Yobe River channel is deeper under the bridge and acts as a sink, inducing subsurface flow convergence, and creating a pool of water.

Similarly following the 1992 flood recession, the piezometric levels remained above the river level and groundwater commenced contributing water to the river flow. The groundwater level decline was rapid for the period immediately following flood recession and water levels in most piezometers fell by more than one metre below their peak levels in the first month of flood recession (figure 6.2c and 6.3). The water level decline continued, albeit at a reduced rate, throughout the dry period, groundwater level (except in P6) reaching its nadir at the end of the 1993 dry season. By 27/6/93 (day 410 in figure 6.3) the bank storage, i.e the amount of water that went into the aquifer storage during the 1992 flood, was exhausted. The water level decline in all piezometers north of the river was nearly exponential, consistent with the behaviour of river-aquifer system making the transition from peak flow to baseflow (figure 6.3). Also the synchronised water level rise and fall in piezometers A to G suggest that there were few subsurface discontinuities along this transect and the aquifer may be considered to be homogeneous.

The exponential groundwater level recession pattern was also displayed by piezometric levels south of the river, except for water level in P6 (figure 6.3b). Water level in P6 was declining following flood recession but rose suddenly by about half a metre and remained stabilised at an elevation of 333.0m above sea level throughout the remaining period of recession (figure 6.3b). A cross section (figure 6.2c) depicting the positions of the aquifer water level following the 1992 flood recession shows that water level in P6 fell just below the upper clay unit and remained at that elevation without returning to its 1992 dry season position. The reason for this anomalous behaviour is discussed later.

The groundwater level rise and subsequent fall during the wet and dry periods respectively shows the degree of interconnection between the Yobe River and the adjacent alluvial deposits. It also indicates that lateral seepage of river water into the aquifer is one source of groundwater input in addition to other possible aquifer inputs such as infiltration from ponded flood water. Also the return of the water levels in most piezometers to their dry season positions (1992) following flood recession indicates that under the present condition much of the water that went into the aquifer storage during the wet season emerged as baseflow and the net annual groundwater recharge is limited by the small regional outward gradients observed by IWACO (1985) and others.

In general, three conditions are necessary for significant bank storage and baseflow to occur (Kondolf et al, 1987):

1) the river reach must be subject to river stage changes from passage of flood flows;
2) the bank material must have a high hydraulic conductivity; and
3) there must be a sufficient volume of permeable bank material to provide significant storage relative to river flow.

Condition (1) implies that downstream reaches are generally more favourable to bank storage and baseflow than are headwater reaches because their greater drainage areas can produce larger flood peaks and thus greater changes in river stage. However in the Yobe basin the downstream reaches are inadvertently being flanked by more alluvium (fine sediments) than bedrock dominated headwater reaches; thus the amount of bank storage and subsequently baseflow which would have been created by condition (1) would be reduced significantly by condition (3). Pinder and Sauer (1971) developed a numerical model for bank recharge which coupled dynamic equations for one dimensional open channel flow and two dimensional transient groundwater flow. They found the importance of bank storage to depend most heavily on the hydraulic conductivity of bank materials and much less on the width of the alluvial aquifer. The other feature of the Yobe basin is that it is a losing system; flood discharges and durations reduce in the passage of flow downstream, resulting in less favourable conditions for recharging bank storage.

Condition (2) is the most significant constraint, one that prevents bank storage/baseflow from being an ubiquitous phenomenon in the Yobe valley. In alluvial deposits, high hydraulic conductivities characterise channel sands and gravels but not overbank muds. Thus high gradient, straight to braided rivers of the upper Yobe, dominated as they are by sand and gravel, would be more likely to produce alluvial fill of high hydraulic conductivity than would low gradient, meandering rivers of the downstream in which the overbank muds often constitute the largest part of the sedimentary facies. Thus as the hydraulic conductivity of the Yobe alluvium decreases downstream as a result of the sediment becoming finer and/or the localised reduction in aquifer conductivity due to increase in clay content, the river-aquifer system would be expected to yield lower magnitudes of bank storage and baseflow than those that could be obtained upstream.

Factors Influencing Yobe River-Aquifer Interactions

The results presented in the foregoing subsections highlighted the nature of the Yobe River-Aquifer interactions. An interesting facet of the observed interactions is the manner in which the north and south sections of the aquifer responded to the river stage increases and flooding, and subsequent recession. It was observed, for example, that large portions of the north aquifer were responsive to all changes in the river stage, while most parts of the south aquifer experienced appreciable increase in water levels only when the river stage was near its peak and the flood plain surface was mostly inundated. This was partly attributed to the differing initially confined-unconfined conditions associated with the two sections of the aquifer. Also other features are thought to have affected the river-aquifer interaction and are discussed below.

The physical and geologic settings, and the hydrogeologic characteristics of the two sections of the aquifer are contained in table 6.2. The information given in this table can also be seen in figure 6.2. The major differences between the two sections of the aquifer inferred from figure 6.2 and recorded in table 6.2 are:

1) The aquifer is connected directly to the river only on one side. The aquifer is exposed to the river channel on the north side while on the south there is a barrier formed by the upper clay unit;

Table 6.2. Comparison of aquifer characteristics in the north and south sections

Descriptive parameter	North section	South section
Upper clay unit thickness	Thicker at the centre and reduces to 1.0m and 1.50m thick near the river and the north floodplain margin respectively	The clay unit is 4.0m or more thick near the river and the southern margin. It is about 1.0m at the centre
Aquifer thickness	The thickness ranges between 10 and 15m	Thicker at the centre. Thickness ranges between 9 and 16m.
Aquifer material	Composed mainly of sand and gravel	Composed of sand and gravel but three boreholes (P5, P7 and P8) contained clayey sand.
Aquifer contact with the river	The aquifer is exposed at the river bank	The upper clay unit extended down into the river bed and forms a seal on the aquifer.
Dry season aquifer regimes	The aquifer is confined except near the river and the northern margin	The aquifer is mainly unconfined. It is confined only near the river.

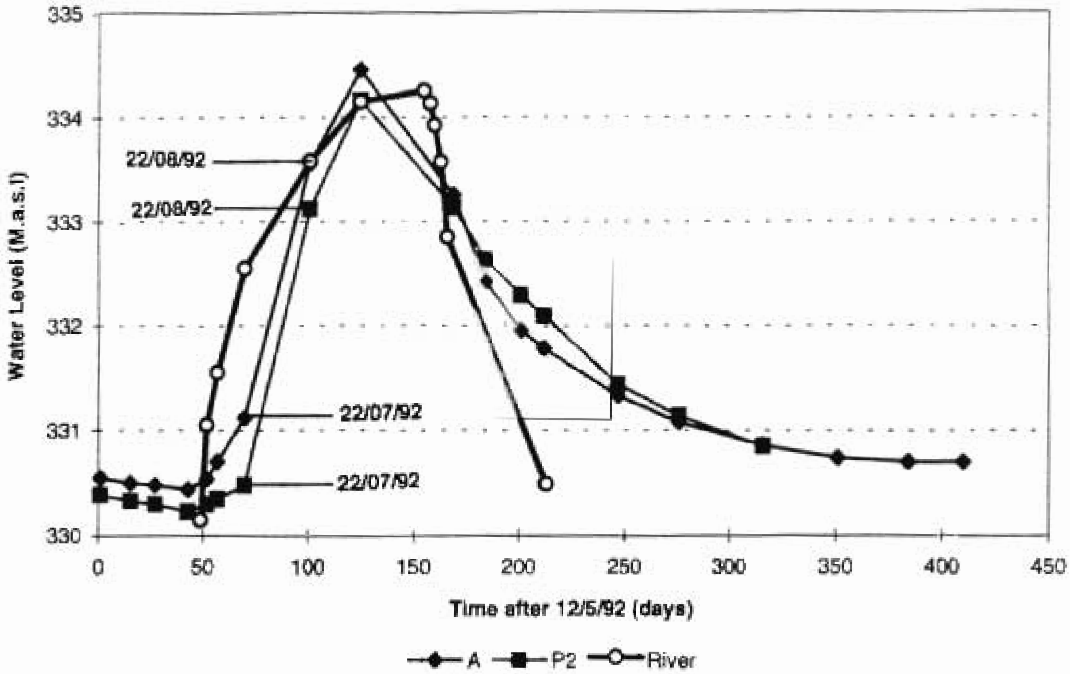
2) the quality of the material comprising the aquifer is not the same on both sides. The aquifer is composed of "clean" sand and gravel on the north side while on the south three of the boreholes (P5, P7 and P8) contained clayey sand ; and

3) the aquifer hydraulic head distribution prior to the commencement of the 1992 river flow is not identical on the north and south sides. Most parts of the aquifer on the north are under confined condition in the dry season while on the south the aquifer is mostly unconfined (figure 6.2a). The upper clay unit maintains a thick, nearly uniform, cover over most of the north aquifer. This unit is thinner (about 1.0m thick) at the centre in the south. The effect of aquifer state on the river-aquifer interaction has been discussed earlier.

***Restriction of flow caused by the presence of less pervious channel*:** In the wet season, as shown earlier, the hydraulic gradient is from the river to the aquifer, and the amount of river water seeping into the aquifer would be influenced by the hydraulic characteristics (permeability) of the material comprising the channel bed and banks. Consequently the seepage condition on the south section, where the aquifer is intruded into near its outcrop in the river by a low permeability clay, is likely to differ from that occurring on the north side.

Figure 6.5a compares hydrographs of piezometers A and P2 which are located at distances of 130m and 140m from the river on the north and south sides respectively. Also included is the 1992 river stage hydrograph. The two piezometers responded in a similar manner to the river stage changes; their levels rose during the wet season, and fell back coincidentally following flood recession. Nevertheless, a close examination of the two hydrographs reveals that P2 response to river stage increase was slightly lower than that in A. For example, by 22/7/92 the river stage increase was 2.40m and the corresponding increase in water level in A was 0.68m which is twice that obtained (0.33m) in P2. This difference is thought to be due to the presence of the "clay wall" at the south river bank which shields the aquifer from the river. The presence of this layer reduced the input pressure into the aquifer by altering the characteristics of the

(a)



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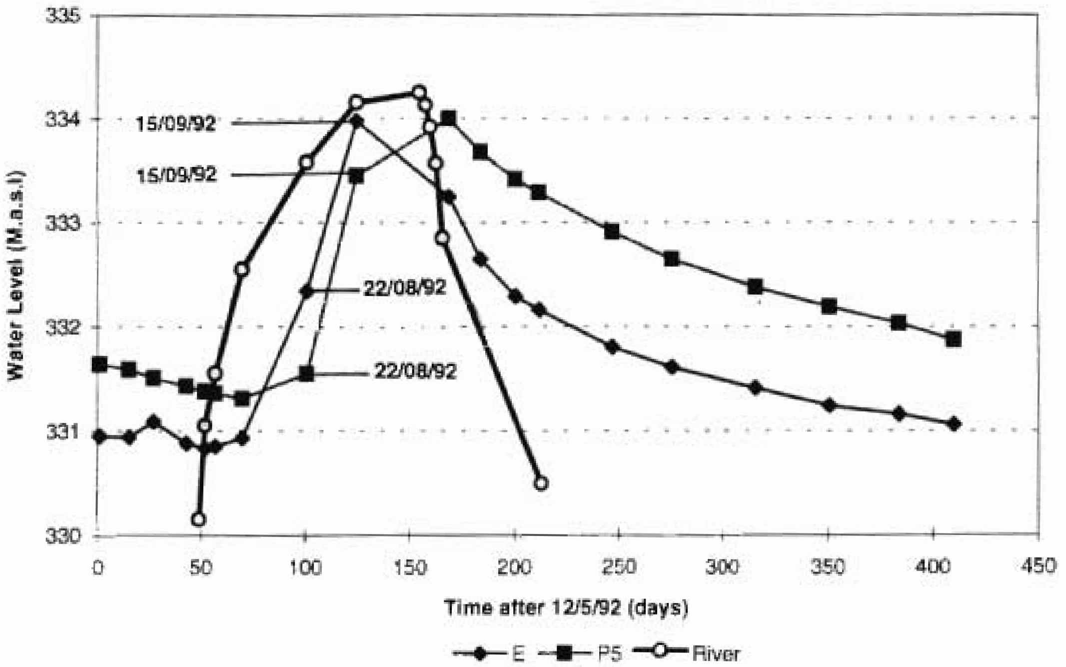


Figure 6.5. Comparison of water level changes in piezometers located at similar positions on the north and south sides of river respectively. (a) piezometers A and P2 located at 140m and 130m on the north and south respectively; (b) piezometers E and P5 located at 1140m and 1136m on the north and south respectively.

(c)

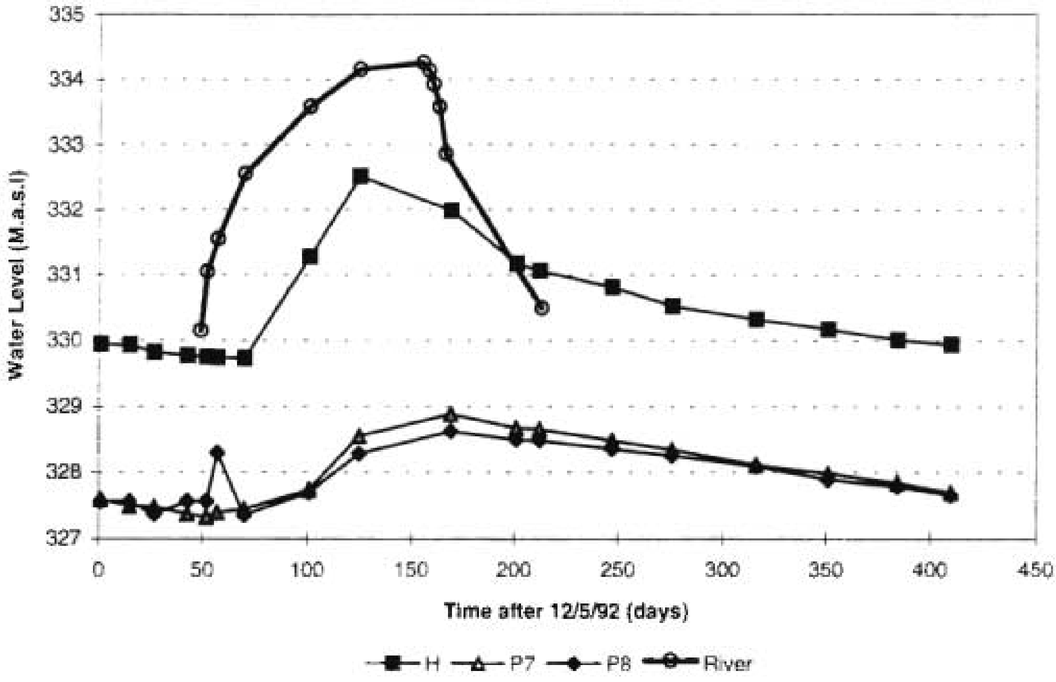


Figure 6.5 cont'd. (c) piezometers H (3100m) and P7 and P8 (2900m) located at the northern and southern margins of the flood plain respectively.

seepage wave and this resulted in lower responses in piezometer P2. This effect is reflected in the rising limb of the P2 hydrograph throughout the flooding period; the measured water level in P2 was always lower than that in A (figure 6.5a). The effect, however, had less influence on P2 during recession; the recession limb of P2 hydrograph is almost identical to that corresponding to A.

The conditions dominating the recession phase of the hydrographs are analogous to those existing in the recovery phase of a pumping test (Rushton and Tomlinson,1979). Conditions during the recovery phase of a pumping test are dominated by the aquifer properties and are only modified close to the well. Consequently the recession of the P2 hydrograph was influenced more by the aquifer properties and less by the properties of the less pervious layer at the river channel face. The aquifer consisted of materials of similar composition at A and P2 locations and water level recession took place at a similar rate and produced the two coinciding exponential curves (figure 6.5a).

Effects of variation in aquifer material composition: Figure 6.5b shows water level changes in piezometers E and P5 which are located at distances of 1140m and 1136m respectively on the north and south sides of the river. Water levels in both piezometers were under confined condition prior to the commencement of the 1992 river flow (figure 6.2a). But despite the two being under similar distance and hydraulic condition, increases in water level in P5 due to the river stage rise was always delayed and lagged behind water level increase observed in piezometer E during the same period. Water level in P5 was perched at 331.50masl and stood above that in E (day number 0 to 70) in the dry season. It practically remained at this position throughout July and August 1992 and only increased after a substantial rise in river stage had occurred (15/9/92). Piezometer E on the other hand was responsive to the earlier increases in the river stage during which its water level rose from 331 masl to higher than 332 masl.

The delayed response is also reflected in the recession limbs of the two hydrographs; water level in P5 always remained above that in E during flood recession.

The differing response from E and P5 was the result of a high clay content observed at piezometer P5 location during drilling combined with effect from the bank wall discussed above. During the installation of P5 a sequence of clayey materials, clayey sand and clay, was encountered while at borehole E the deposits encountered were mainly "clean" sand and gravel. The effect of increased clay content at P5 was to reduce aquifer transmissivity and this has the inevitable consequence of retarding water movement into or out of the aquifer around this piezometer. Hence it took a longer time for water to flow into, and subsequently drain out of, P5 than in other parts of the aquifer such as E with less clayey materials.

A more pronounced increase in clay content leading to further reduction in the transmissivity exacerbates the delayed response and drainage phenomenon, and this was the case near the southern margins of the floodplain where the clay content was higher (figure 6.5c). Figure 6.5c compares water levels in piezometers H, P7 and P8 located at the northern and southern margins of the floodplain respectively. As can be seen in this figure the water level changes in P7 and P8 in response to river stage changes were small compared to that in H. The lack of significant water level changes in P7 and P8 suggest that there was little recharge and discharge at these locations during the recorded period. Low discharge in an area of extremely high transmissivity and storativity could also produce little change in water levels. However, considering the types of sediments present at these piezometer locations- an interbedded sequence of clayey sands and clays (see figure 6.2)- it is obvious that transmissivity is lower than transmissivity elsewhere in the basin. A balanced recharge-discharge condition could produce long term stability but, annually these piezometers might have been expected to have a similar magnitude of water level changes as in piezometer, H, which was located at a similar distance from the river, and existed under an unconfined aquifer condition prior to the commencement of the 1992 river flow, as P7 and P8. The transmissivity reduction here had caused localised steepening of the hydraulic gradient towards the margin of the floodplain (figure 6.2). This observation seems to confirm our earlier assertion that little outward groundwater flow was taking place at this boundary towards the adjacent upland.

Flooding effects

All the above comments were based on the river stage changes and interpretation of the observed trends in piezometric levels. Implicit in the above observations, however, are the possible effects of flooding on groundwater levels; most of the floodplain remained inundated for nearly two months, between September and October 1992. Flooding can have an effect on groundwater levels, for as Meinzer (1932) noted: 'The water level in a well is sensitive to every force that acts upon the body of water with which the well communicates'. This is particularly true in the case of confined groundwater and it is, therefore, imperative to highlight the possible effects of flooding on groundwater levels.

In general, flooding has at least two possible effects on groundwater levels:

1) infiltration into the soil to cause a water level increase; and

2) imposition of a load on the underlying formations.

Clearly, however, there is a methodological difficulty in separating aquifer response to river stage changes (lateral recharge) from responses due to the effect of flooding (vertical recharge and loading). Floodplain inundation and maximum river stage occured simultaneously; thus the effects of the two on the groundwater level were linked and could not be easily separated.

Infiltration

Infiltration into the Yobe alluvium occurs following surface inundation in response to the head gradient created by the floodwater at the soil surface. Ideally water will continue to infiltrate into the soil until the surface water supply becomes limited, i.e the head gradient drops to zero. However, soil infiltrability and its variation with time depend not only on the initial soil wetness and suction, but also on other intricately interacting factors such as the texture, structure, and uniformity (or layering sequence) of the soil profile. These latter factors in most cases influence the amount of infiltration that takes place over a given time period (Hillel,1971).

The soils, clays and silty clays, comprising the upper layer of the floodplain have swelling and cracking properties which are likely to be important hydrologically. Twenty (20) profile pits were dug to observe crack depth and soil profile. Crack depths were also estimated by inserting a wire into several cracks. During the dry season cracks developed and extended to a variable depth: most crack dips are vertical or near vertical and the clay becomes increasingly moist with depth below ground surface. The observed crack depths ranges between 30cm and 40cm below the ground surface, and below this depth soil moisture content increases significantly within a few centimetres and visible cracks diminish.

The foregoing observations have two possible hydrologic implications:

1) During the wet season the cracks would provide a "short circuit" route for rain and flood waters enabling the soil to re-wet quickly; and
2) with clay becoming less cracked and wetter downward, the bulk density is likely to increase with depth because of a decline in available capillary pore space and in hydraulic conductivity. Thus the lower part of the upper clay unit might be acting particularly in the initial stages of flooding as a sort of "impeding" layer to vertical flow.

It is recognised that the physical characterisation of the clay given above is limited in scope and cannot give a firm definition of the hydraulic characteristics of the clay layer. Thus the hydrological characteristics of the clay layer can only be regarded as the most likely, based on the present observations, until further investigation is undertaken.

Infiltration Concepts in Cracked Soils

The process of infiltration into cracking soils under ponding in similar geological settings and climatic conditions as ours has been described by Grismer and Tod (1991). The study was carried out in an arid environment (annual rainfall is about 75mm) at the Imperial Valley, United States. Grismer and Tod (1991) noted: "During ponding, large surface cracks fill very rapidly with water and the soil adjacent to the cracks becomes quickly saturated. The network of cracks within the top layer is sufficient to saturate the soil throughout the top layer soon after the arrival of the wetting front. During this wetting process, the surface layer of the soil disperses and its water content becomes higher than the water content of the soil below the surface. Following initial filling of cracks by water near the wetting front, subsequent infiltration of water into the soil is limited by the low permeability of the soil peds, making diffusion of water from the cracks into the soil peds very slow. Even though the cracks do not close completely, the presence of the perched water table limits further downward movement of water through the cracks".

Because the soils in the Yobe floodplain are mostly cracked, it might be expected that the soils under ponding would display hydrologic characteristics similar to those observed by Grismer and Tod (1991). That is, following annual flooding water would flow into the cracks and subsequently develop a perched water table within the upper clay layer. In this case a point of consideration is the fate of the perched water table; does it remain standing within the clay throughout, or is it gradually dissipated by seepage downward into the underlying aquifer or by lateral water movement at shallow depth within the clay itself towards possible discharge areas?.

The occurrence of lateral flow within the clay in the present study site is likely to be a complex process and was possibly non-existent during the wet season since all possible discharge areas (depressions) were flooded, and perhaps of limited magnitude at other times because the topographical gradient of the site is far too gentle to induce rapid lateral water movement. It is certainly thought that the use of the term "interflow" is inappropriate here and perhaps at best, if this movement exists, it would be more appropriately described as lateral "clay" seepage. Clay seepage might possibly emanate from the relatively high sections (ridges) of the flood plain and emerge as diffuse seepage at the shallow depressions (swales) during the dry season. Diffuse seepage in fractured clay governed by subtle and small-scale topographic irregularies caused by ridges and swales have been observed by Keller et al (1986) in some North American flood plains. The flow was due to the lateral interconnection of fractures and retarded vertical flow caused by an underlying low permeability clay. Such flow was limited to the upper parts of the clay where cracks were concentrated and laterally connected to one another. Cracks extending downward were also capable of transmitting water but lateral interconnection was limited and groundwater movement at depth was predominantly vertical in response to the seasonal changes in hydraulic gradient.

The dissipation of a perched water table in soils of low permeability by vertical seepage has been observed by Kidder and Lytle (1946) during a groundwater investigation in heavy glacial till soils of north-east Illinois, USA. They interpreted their well measurements as indicating "...the level of the top of the retreating water surface in the soil"...and continued "...this is not the surface of the permanent water table but instead is the top of a new mass of water that is moving downward through the profile to the water table...". This suggests a perched water body where percolating water has developed a positive head and is continuing to infiltrate slowly downwards.

Drawing from the above, it appears that flow through the upper clay unit of the Yobe basin alluvium into the underlying sand and gravel aquifer may take place during flooding and persist even after the surface water supply has been removed. Thus it is possible that the observed exponential decline of the well hydrographs (figure 6.3) could be a function of lateral drainage through the aquifer combined with a much slower, but continuing, vertical water movement through the clay into the aquifer. In addition the later decline in the recession rate is probably a reflection of the progressive decrease in clay hydraulic conductivity with depth as well as changes in vertical hydraulic gradient.

Infiltration Magnitude in the Present Site

As has been stated earlier, and from the foregoing considerations, it is difficult to explicitly isolate the effects of the surface infiltration recharge on the aquifer water level from those due to lateral channel seepage. Nonetheless, the upper clay unit has variable thickness and comprises materials of variable texture and composition and it is likely that the hydraulic conductivity of this unit would also vary within the present site (Table 6.3). The hydraulic conductivity values for similar units deduced from regional studies and given in table 6.3 range from $9.4.10^{-7}$ to 2.010^{-2} m/d. It is likely that similar range of values may be obtained in the present site which would, thus, result in variable darcy fluxes. For the range of values given in table 6.3, for instance, the fluxes under a unit vertical hydraulic gradient would vary from $9.4.10^{-7}$ to 2.10^{-2} m/d. Consequently hydrograph(s) from a piezometer(s) installed at a location where infiltration recharge has been significant is likely to have differing characteristics from those where such effect was less significant.

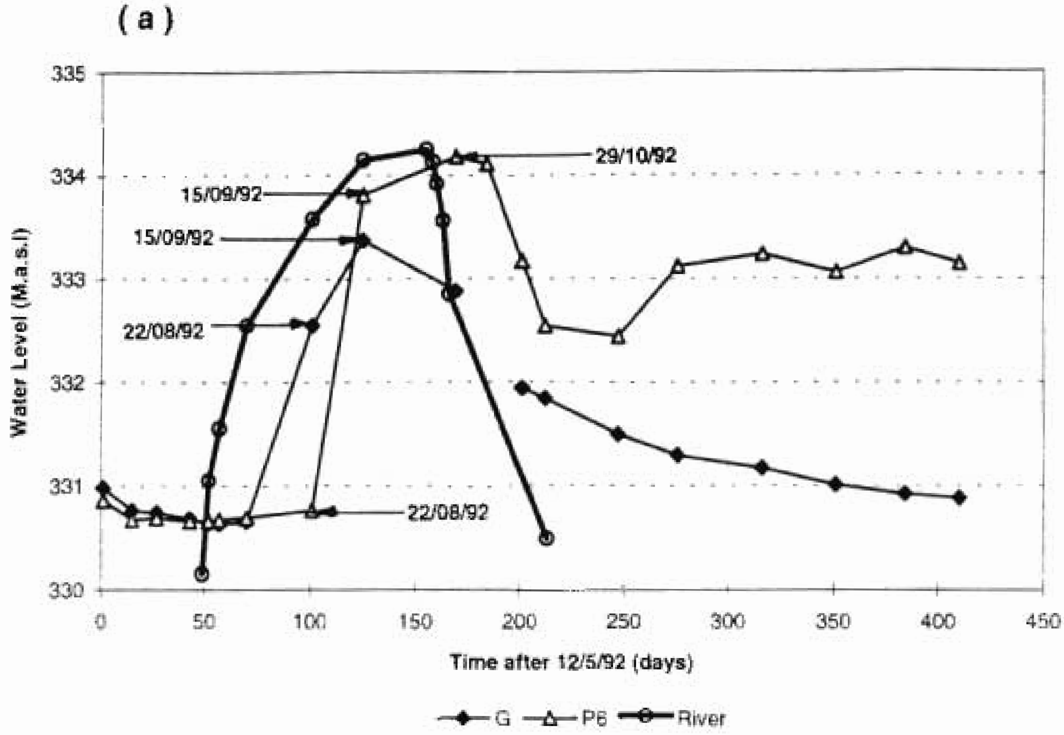
Table 6.3. Hydraulic conductivities of the argillaceous part of the Yobe alluvium deduced from regional studies.

Source	Data points	Hydraulic conductivity (m/d)
Diyam (1979)	13	9.4×10^{-7} to 2×10^{-2}
Haskoning (1979)	3	3.5×10^{-6} to 1.7×10^{-5}
N.A.P.C. (1979)	6	8.6×10^{-6} to 3.5×10^{-3}

Note: All these tests were conducted upstream of the present site.

Figure 6.6a compares hydrographs of piezometers P6 and G which are located on the south and north sides of the channel respectively. Both piezometers are located at a similar distance, 2100m and 1900m respectively, from the river. This comparison was made on the presumption that the two hydrographs would exhibit similar features in response to the river flow and flooding since the two are located at similar distances from the river. It is clear, however, that the response was delayed in P6, and the shape and amplitude of its hydrograph are also significantly different from those corresponding to G. The total increment by 15/9/92 was 2.70m in G and seventy percent (70%)(1.87m) of this took place between 1/7/92 and 22/8/92 (53 days). All this time elapsed and water level in P6 increased by only 0.10m. Piezometer P6 was responsive to the events after 22/8/92 when river stage rise was followed by flooding. The time between the occurence of this event and piezometer P6 responding was short as indicated by the steep rising limb of its hydrograph (figure 6.6a). Water level in P6 rose by more than 3m in 24 days and continued rising until its apogee was attained (334.18masl) on 29/10/92 (day number 169), two weeks into the 1992 flood recession. From then on its level declined and dropped to 333.17masl (1m lower than its peak) by 30/11/92 (day number 201), and remained at this elevation throughout the remaining recession period. This provided a striking contrast with the very short period it took for the water level to rise by more than 3m during flooding, and also the rapid fall in level observed in piezometers G and D (figure 6.6a and b). Water levels in these piezometers fell back rapidly as the river receded and continued declining throughout the dry season. The rising limb of the D hydrograph is also similar to that of G, i.e more than 65% of its level was attained before 22/8/92.

The hydrograph of P6 was also compared with those from P4 and P5 which are located at about 800m and 1136m respectively from the river along the same (south) transect. The total level rise in P4 water level by 15/9/92 was about 3m of which more than 50% (1.79m) was attained before 22/8/92 (figure 6.6c). The rise in P5 occurred mainly after 22/8/92 and is characterised by a steep rising limb which coincides with



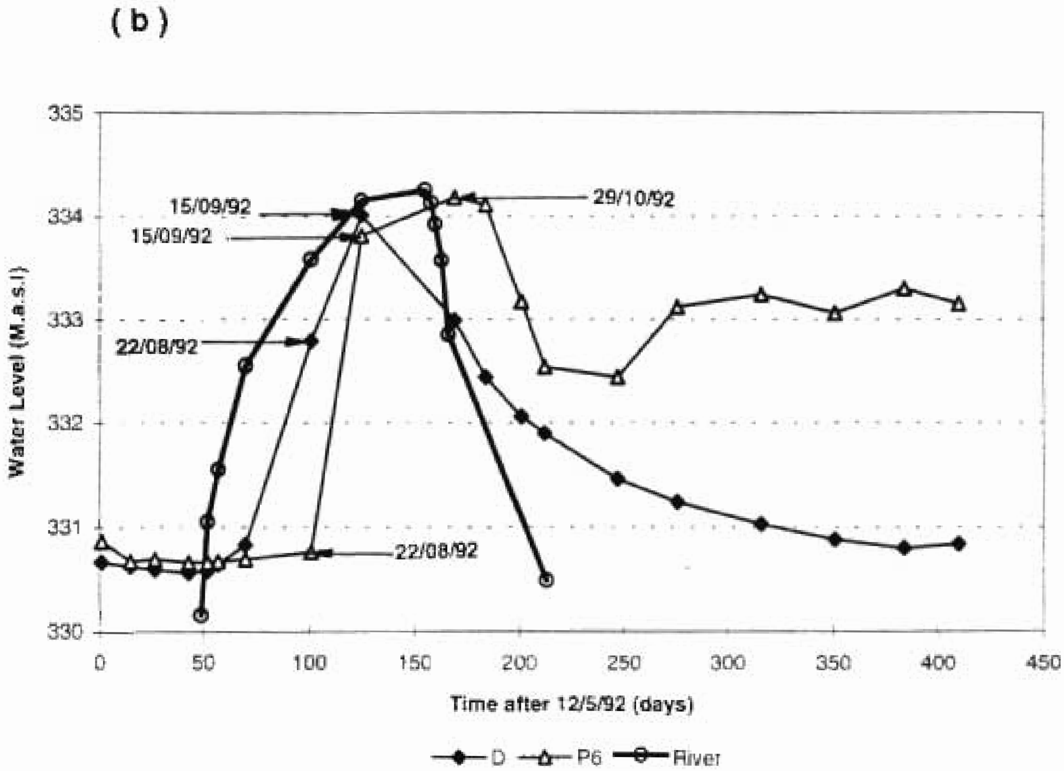
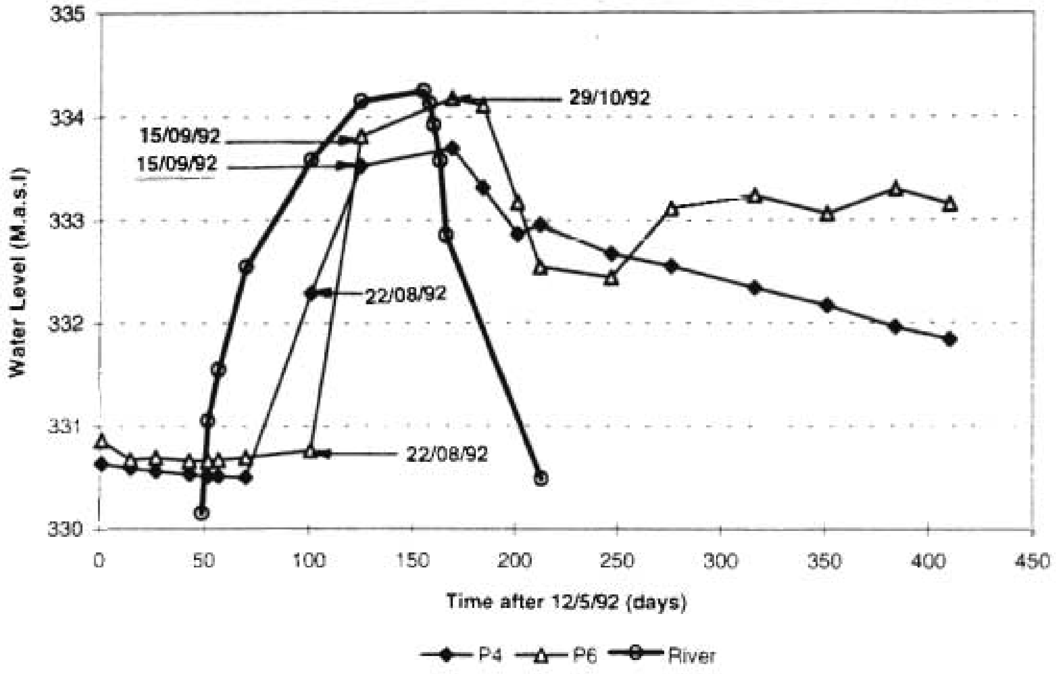


Figure 6.6. Hydrographs depicting vertical infiltration effects on aquifer water levels. (a) piezometers P6 and G located at 2100m and 1900m on the south and north sections respectively; (b) piezometers P6 and D located at 2100m and 900m on the south and north sections respectively.

(c)



(d)

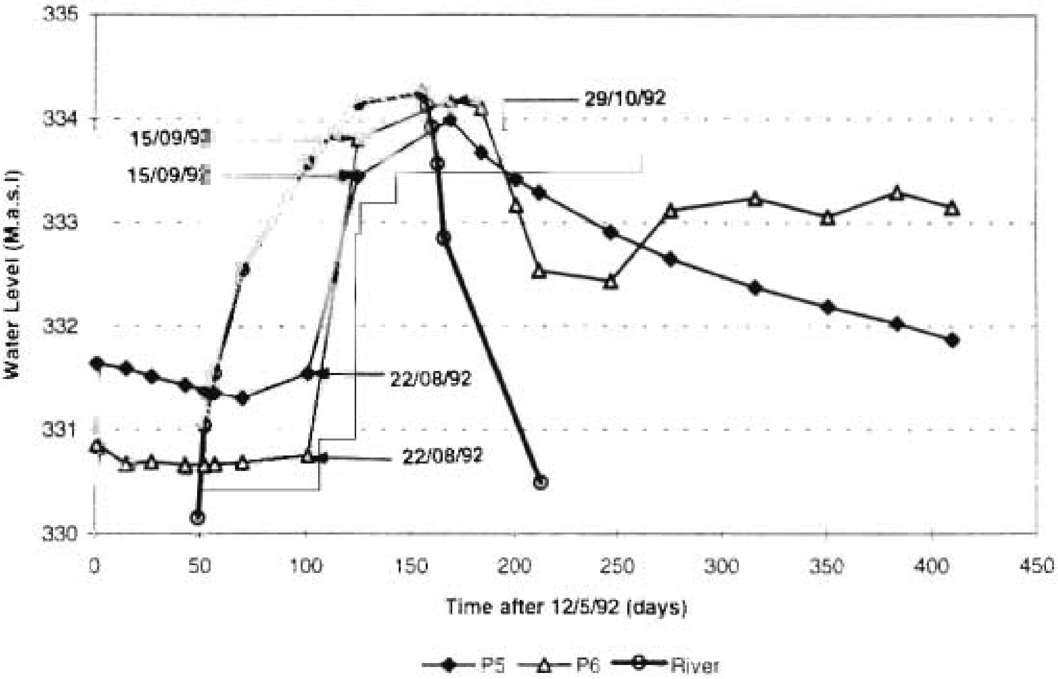


Figure 6.6. cont'd. (c) piezometers P6 and P4 located at 2100m and 800m along the same transect in the south section; (d) piezometers P6 and P5 located at 2100m and 1140m along the same transect in the south section.

that of P6 (figure 6.6d). The total increase in P5 water level was 2m by 15/9/92, less than that in P6 by 1m.

The foregoing observations, and the discussions given throughout this chapter, show that the water level increase in P6 was negligible when the river was confined to its channel. It quickly became larger than the levels in P4 and P5 following inundation of the flood plain in September 1992. It also became larger than the total increase in G which is located at a similar distance as it from the river on the north side as a result of this event. In addition, the water level in P6 stood above levels in all the 20 piezometers observed in the present study after cessation of the 1992 flood (figure 6.3 and appendix B).

The water level increase in P6 was not expected to be so large if the response was due mainly to input from channel seepage. Piezometer P6 is located farther (2100m) from the river than most and the channel seepage influence was always seen to decrease with distance inland. This decrease was further accentuated by the retarded (and slower) lateral flow towards the south induced by the presence of an initially unconfined region in the south centre, and the existence of less pervious river bank and aquifer materials. These effects practically restricted P6 from responding to the earlier channel seepage that resulted from a 3.40m river stage increase between 1/7/92 and 22/8/92 (53 days). This aspect has been discussed earlier. Because of these it would appear inconceivable that the observed water level rise of more than 3m between 22/8/92 and 15/9/92 (24 days) would be entirely due to seepage input. If this was so, such a rise would equally be reflected, and the magnitude would be even larger, in those piezometers which are located closer to the river. Thus this anomaly can only be accounted for if there was an additional water input into the aquifer that resulted from an alternative recharge mechanism.

We attribute these apparent differences and the further rise in the P6 hydrograph at the time when (a) no rainfall occurred, (b) the river stage had fallen, (c) the flood plain surface was drained, and (d) the lateral hydraulic gradient was towards the river, to an additional input from infiltration recharge. The magnitude of such recharge invariably is attenuated at points where the upper unit is thick since the infiltration pulse would take a long time to reach the aquifer water level. Infiltration recharge was significant at P6 possibly because the clay unit is thinner adjacent to P6 than at the other locations considered. This unit is about 1m thick in the P6 vicinity and was likely to have provided a shorter flow length to the infiltrating water than, for instance, in the G surroundings where it is about 4m thick (figure 6.7). It is possible that this together with a favourable (clay unit) hydraulic conductivity provided greater opportunity for flood water to infiltrate and move downward into the aquifer.

The rate of vertical infiltration can be estimated roughly from Darcy's law. For instance for the given hydraulic conductivity of 2.10^{-2} m/d (table 6.3), the wetting front would travel a vertical distance of 1.0m in 50 days under a unit vertical hydraulic gradient. The one metre thickness is equivalent to the clay thickness near P6 and the calculated time (50 days) approximates the period from 1/9/92 when surface flooding began, to 29/10/92 when the P6 hydrograph attained its peak. If infiltration at this rate took place, it would account for most of the observed increase in P6 water level. The infiltrating water, however, would require a longer time to reach the aquifer water level at other locations such as G where the clay is thicker. It should be also noted that the hydraulic conductivity used above is at the high end of the range in table 6.3; on textural grounds alone it is judged that most parts of the Yobe floodplain clay hydraulic conductivity is likely to be nearer 10^{-5} or 10^{-6} m/d. Vertical fluxes would be correspondingly low and thus infiltration recharge is likely to be associated with limited sites.

The infiltration recharge effect on the aquifer water level in the north and south sections can be seen in the cross sections shown in figure 6.7. Piezometric levels on three different dates i.e 22/8/92, 15/9/92 and 13/11/92 are shown. The north section is characterised by a rapid fall in water level, while a slower level decline is a feature of the south section. On 13/11/92 the piezometric level in the north has fallen back from

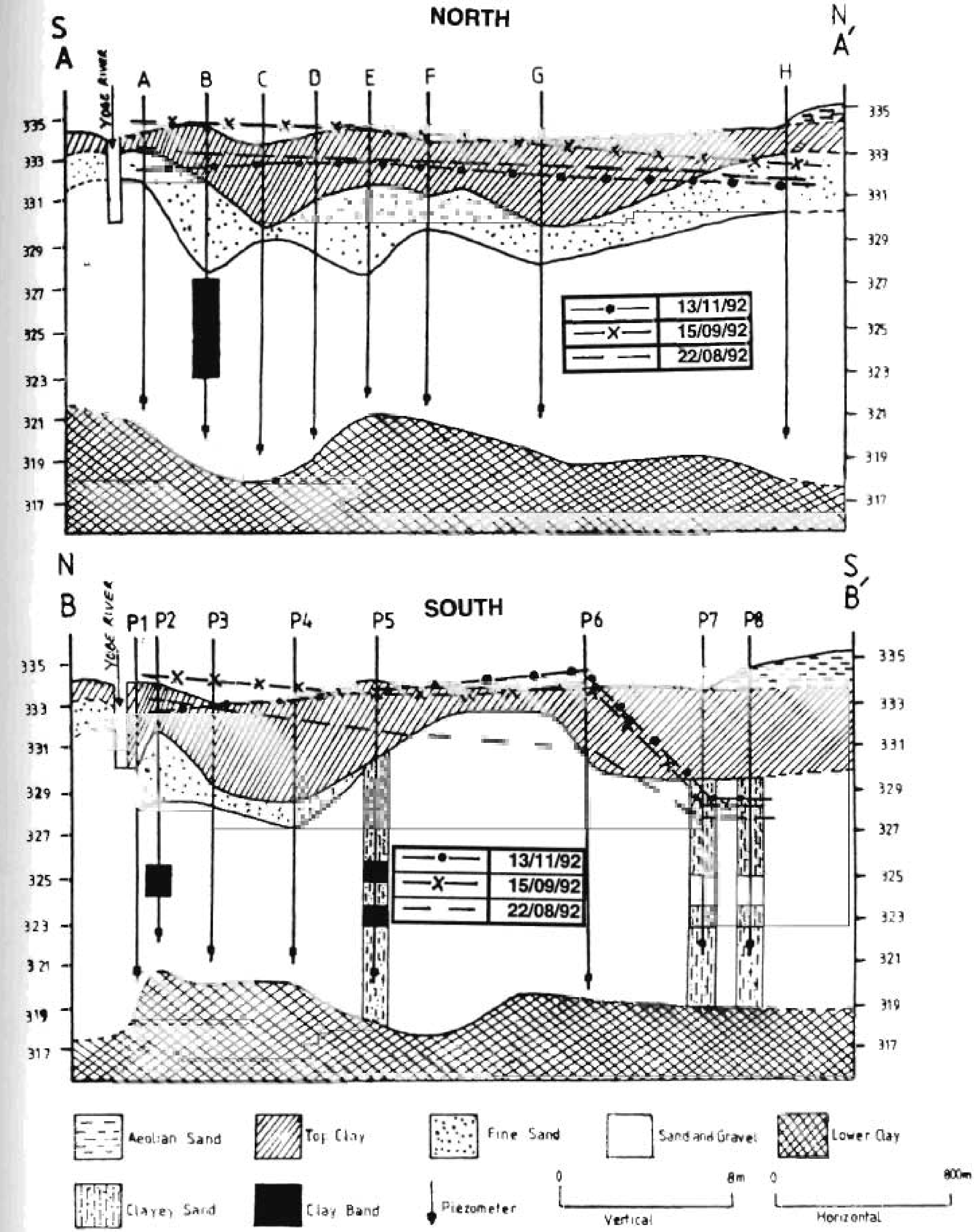


Figure 6.7. The magnitude of infiltration recharge along the north and south transects. Water levels on three dates, ie. 22.8.92, 15.9.92 and 13.11.92 are shown.

a high point on 15/9/92 and returned close to its positions on 22/8/92. The recorded aquifer water level on 22/8/92 was due to input from lateral channel seepage alone and the return of the piezometric level close to it immediately following cessation of river flow and flooding can either be due to: 1) lateral redistribution of significant volumes of the recharged water or, 2) pressure dissipation as a result of a confined aquifer response. Given the small lateral hydraulic gradients (4.10^{-4} to 6.10^{-4}) prevailing in the site though, it seems highly unlikely that recharge redistribution would effect such a rapid decline within a month of flood recession. It is likely, however, that the rapid decline was a confined aquifer response in which the excess pore pressure was dissipated as a result of load release at the river channel boundary. As the river stage receded, the load on the aquifer was removed, and the rapid water level increase observed along this transect in the wet season was reversed by an equally rapid decline. Little change in the stored volume took place.

In contrast in the south central region water level on 13/11/92 was slightly above that on 15/9/92. This was so because parts of this section of the aquifer appears to have had greater infiltration recharge. This greater opportunity for localised recharge resulted in the water level high, the effect of which persisted throughout the year and was never dissipated completely even by 27/6/93, as eveidenced in the maps and cross sections given earlier (figures 6.1 and 6.2). Conversely, the lack of opportunity for significant vertical recharge, together with a confined aquifer response resulted in the absence of a similar condition in the north.

Prospective Sites for Infiltration

Detailed monitoring of water level hydrographs indicated variations in infiltration recharge within the study site. Such recharge was more in the central south region than in the north. This observation seems to suggests that flood water infiltration in the Yobe valley is likely to be associated with limited sites rather than one which is widely distributed spatially. This highlights the need to further investigate the reaction of the groundwater system to short lived, local, inputs of water, compared to recharge more uniformly distributed spatially. This is required to assess the relative significance of localised vertical recharge and channel seepage. In particular, aquifer properties will interact with the spatial distribution and intensity of localised recharge to determine the magnitude of the groundwater mound, and the time it takes to decay and propagate to the rest of the aquifer system.

As a first guide to identifying and/or determining these, we postulate that sites with clay thickness of about 1m or less are potential candidates for consideration. The area covered by this thickness of the clay in the present site is shown in figure 6.8. This map is deduced from the clay thickness distribution map given earlier in chapter five. The shaded portions of this map are the infiltration sites based on the present discussions and occupy about 20% (2.5 km^2) of the approximately 12.0 km^2 study site surface.

Loading

Water level changes as a result of surface loading effects are known to occur in virtually all natural formations. The most easily observed and common examples of such loads are barometric pressure changes and ocean tides. It is a well known fact that these loads result in groundwater level changes, even in deep confined aquifers (e.g Jacob,1940).

Other loading effects, in particular loading due to annual changes of total moisture accumulations above an aquifer have recently been shown to contribute to the annual fluctuations of piezometric levels in aquifers (Van der Kamp and Maathius,1991; Bardsley and Campbell,1994). Simple physical considerations by these authors suggest that such loading effects occur for almost any confined aquifer, but their effects in most cases are obscured by other causes of annual changes of groundwater levels and are difficult to recognise.

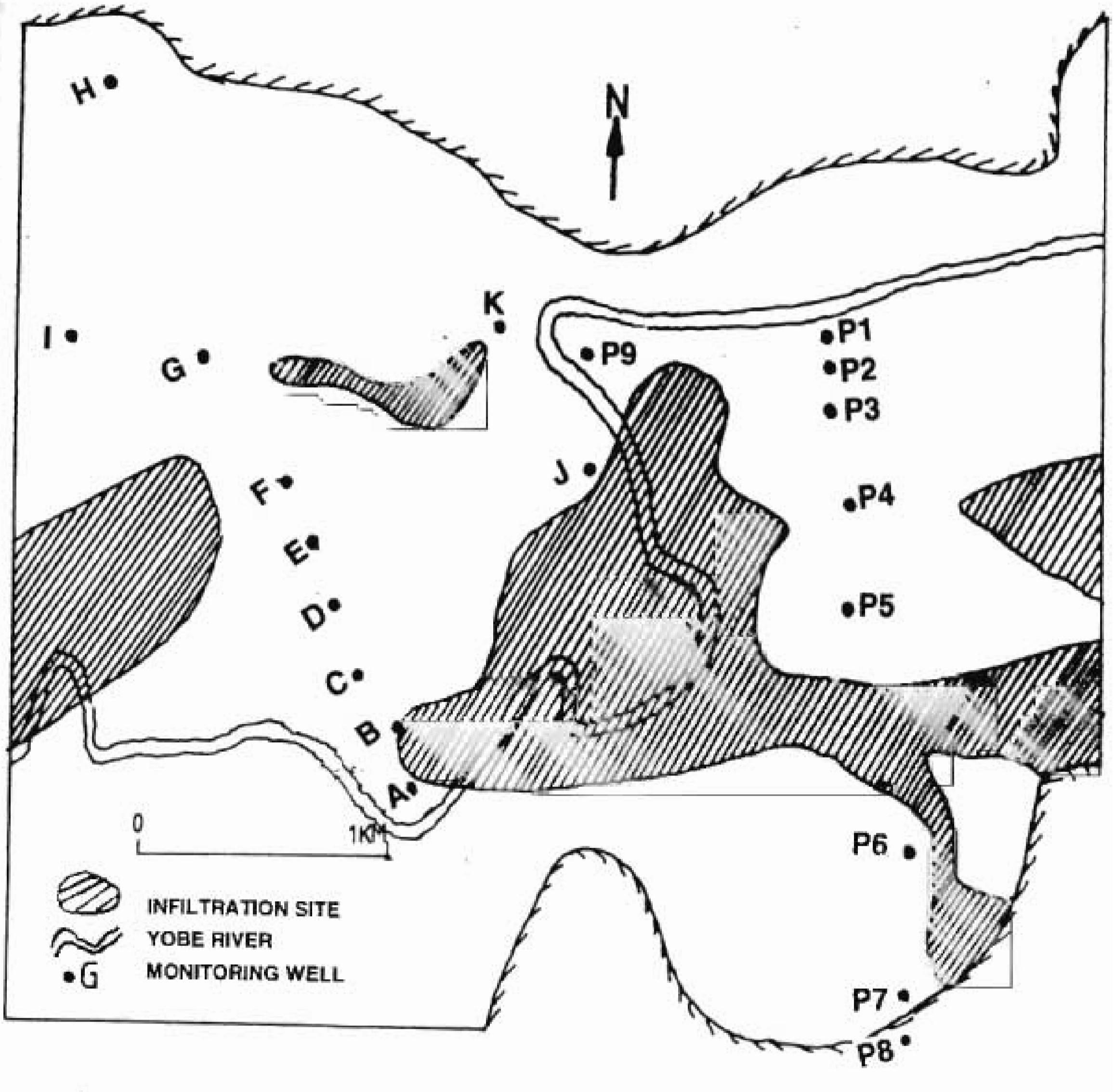


Figure 6.8. A map showing prospective sites for vertical flood water infiltration within the study site. The shaded portions are locations where the surface clay thickness is within 1.0m.

All these fluctuations are due to the internal propagation of stress through the formations and need not involve any flow.

The loading of an aquifer by surface inundation is analogous to loading due to tidal flooding, the difference between the two being the frequency of occurrence; floodplain inundation in the Yobe Basin occurs annually while tidal flooding may occur over a diurnal period. Where a confined aquifer extends beneath a sea, or is inundated as in the Yobe valley, the increasing weight of water associated with the incoming flood/tides imposes a load which the overlying aquitard may not be able completely to resist. In this case, that part of the increased pressure which is not resisted by the aquitard is transmitted into the aquifer and causes an increase in piezometric level in a well tapping the aquifer (Ward,1975). Application of the load compresses the aquifer and induces the greatest pore pressure changes in the adjoining aquitards and in the more compressible parts of the aquifer. The resultant transient hydraulic gradients lead to flow within the aquifer and the aquitards, and the net result is a complex pattern of transient flows induced by the transient external loads (Van der Kamp and Maathius,1991). Subsequently, when the load is removed, the pressure drops to a minimum and then recovers toward its initial value.

Since the aquifer in our study site is alluvial, at most locations confined, and subject to annual flooding, it is logical to think that the measured piezometric levels also include some effects due to loading.

Comments and Conclusions

The study of groundwater flow at the study site not only reveals features typical of an interconnected river-aquifer system in an arid environment but allows for the examination of these characteristics under a range of river stages.

The hydraulic head in the alluvial aquifer is governed by changes in river stage. This is revealed by a groundwater level increase during the wet season and a subsequent decline following river level recession. Observations indicated that during sustained high river stage the water level increase occurred rapidly throughout the study site and the piezometric water levels rose and remained close (in some cases above the ground surface) to the floodplain surface. Following flood recession, the groundwater level fell rapidly during the first month and continued falling, but at a reduced rate, for several months reaching its nadir by June 1993. In 1992, a wet year, bank storage contributions to river flow were detected eight months after flood recession. On the return of the next wet season and river flow, water flows into the aquifer depending on the amount of water removed from the aquifer during the dry period, and the cycle of storage and recession begins again.

The alluvial groundwater flow pattern was determined by examining the configuration of the groundwater system from water level contours and cross-sections constructed for several dates of the year. Detailed inspections of these maps and cross-sections revealed that the groundwater flow pattern represents a continuum of constantly changing conditions and adjustments. The distribution of the hydraulic head and flow were determined by the rate and areal distribution of recharge, and the geometry and hydraulic properties of the aquifer and the ubiquitous clay cover.

Groundwater flow during the dry season varies slightly from the commonly observed phenomenon in arid climate hydrology where the generalised pattern of flow within the shallow groundwater is controlled by the geometry of the river channel network, with groundwater moving from uplands towards the river channel. Observations on the north section of the study site revealed that groundwater flows from the centre toward the northern margin of the floodplain. Water level in the central region of the aquifer was slightly higher than in the areas near the north flood plain margin and near the river channel and remained confined throughout, and was being steadily drained towards the river channel and the adjacent upland. As a result three distinct groundwater regimes were developed i.e the confined centre and two unconfined regions, one near the river and the other at the flood plain margin. The central region on the south side equally had elevated but unconfined water level and was being drained mainly towards the river channel. Flow towards the south margin was restricted because of reduced transmissivity there induced by high clay content in the aquifer material. The south sector was confined only near the river and the remaining parts of the aquifer were mainly unconfined.

Subsurface flow adjacent to the river may have different flow directions at different locations along the valley, depending on whether or not the aquifer has lateral continuity with the adjacent uplands. The continuity or otherwise between alluvial and sub-interfluve aquifers therefore needs to be considered when applying models to the subsurface flow within the Yobe basin, and when monitoring groundwater and interpreting its water quality characteristics.

The dry season flow pattern was altered when the river stage increased following the 1992 flood. The gradient towards the river was reversed at the early stages of the river flow and from then on flow was directed from the river into the aquifer. A NE-SW trending groundwater mound developed at the centre of the site from which water was being directed towards the NW and SE directions. As a result of this and subsequent infiltration most of the unconfined regions of the aquifer were progressively transformed into confined groundwater.

The filling of the unconfined areas was rapid on the north side and was due mainly to the input from channel seepage. An inspection of groundwater level and river stage hydrographs revealed a rapid transfer of incident river flow and nearly the whole of this section of the aquifer was confined within seven weeks of the commencement of flow. The response was slow on the south side to this (early) input and water levels in most parts of the unconfined sector of the aquifer rose by only a small amount and practically remained at their dry season position. Three weeks later the river stage was near its peak, the flood plain was inundated, and there was a general increase in the aquifer water levels. As a result the unconfined regions of the south switched to a confined state. This conversion was enhanced by input from localised vertical flood water infiltration.

The aquifer water level fell back as the river receded and the transformed sections of the aquifer began the return journey to their unconfined state. It has been seen that this journey was rapid for the whole of the north section and the temporarily confined sections became unconfined in a short period of time. The significant effect of the rapid recession was that little benefit was gained in terms of storage from the large 1992 flood. It took a longer time (5 months) for the water level in the south central region to fall (just) below the base of the cover clay. From then on the recession was even slower and water level in this region never regained its 1992 dry season (25/6/92) position. This sector of the aquifer essentially exhibited features akin to an unconfined aquifer while the north sector behaved practically as a confined aquifer.

The foregoing observations indicate that the two sections of the Yobe alluvium do not only differ in terms of the materials comprising them, but equally are asymmetrical with regards to their storativities. It is therefore necessary to consider the alluvial groundwater system as consisting of these two regimes when designing a mathematical model that would be used to simulate its behaviour, since the storage in the different regions of the aquifer are of significantly different magnitudes. The unconfined regions have high storage coefficients, several orders of magnitude larger than in the confined areas.

The Yobe River is in hydraulic continuity with the adjacent alluvial aquifer, and changes in the river water level affect the aquifer water levels. The principal observation from the study was the variability of the response of the north and south sections of the aquifer to the annual river flow followed by flooding and recession. This variability was manifested spatially over the study site by localised vertical infiltration of flood water into the underlying aquifer, by the non-uniform distribution of aquifer hydraulic head and regimes, and retarded flow at some points within the aquifer. All of the above findings have important implications regarding the development of a necessary model that would be designed to represent the Yobe River-alluvial aquifer system.

6.3 Conceptual Model of the Yobe River-alluvial aquifer System.

The proposed conceptual model of the Yobe River-alluvial groundwater system in the study site is shown by the generalised diagram in figure 6.9. General system characteristics shown in the diagram occur in the site and are the dominant influences on the occurrence and flow of groundwater.

The Geologic Framework

The geologic framework of the groundwater flow in the Yobe basin alluvium within the present site comprised the three major hydrostratigraphic units identified in chapter 5 and indicated in figure 6.9, namely:

1) An upper clay unit;

2) a sand and gravel layer; and

3) an underlying clay unit.

The nature and disposition of these units conformed to a classical fluvial depositional pattern in a low energy (meandering river) environment and were adequately described by a standard fluvial sediment depositional model.

The upper clay unit consists of poorly sorted clay-rich (clay and silty/sandy clay) sediments, and collectively they form a continuous cover over the flood plain surface. The permeability-thickness product of this unit governs local infiltration rates. This unit has variable thickness, and where it is 1m or less thick, it allows vertical flow into the aquifer to take place. One such site was identified in the south central region of the aquifer (figure 6.9).

The sand and gravel layer comprised the aquifer and is sandwiched between the upper and lower clay units. This unit has variable thickness; it is thicker in troughs (abandoned channels) and thinner at ridges that occupy the surface of the underlying clay. Vertically the aquifer grades from gravel, passing up to sand, which decreases in size upwards; more gravels are encountered in channels than at ridges. Equally the aquifer grades laterally from clean sand and gravel to those intermixed with clay-rich

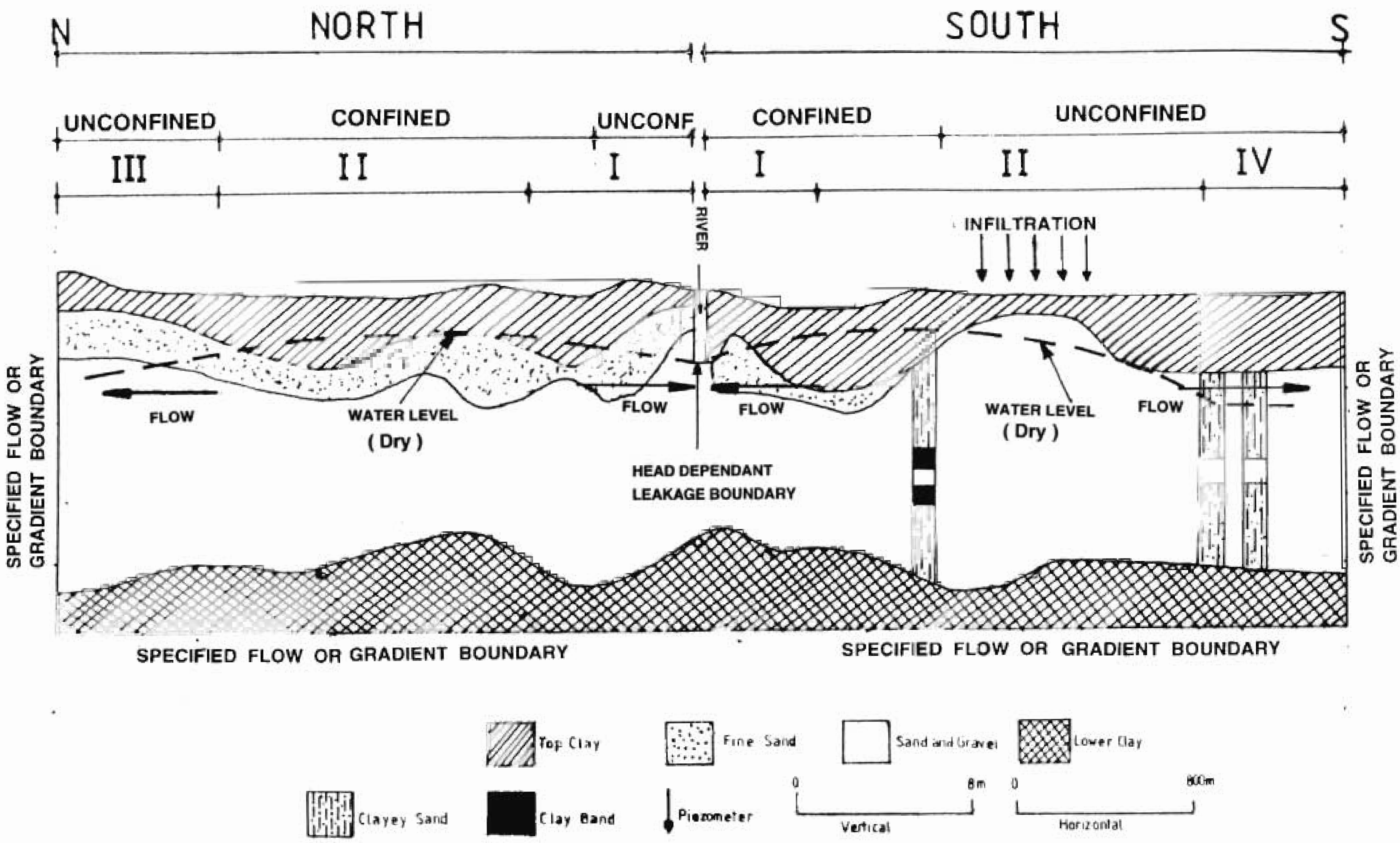


Figure 6.9. The proposed conceptual model of the Yobe River alluvial aquifer system. The identified flow and boundary conditions are as indicated

sediments (sandy clays and clay bands). Such intermixing is more on the south side than on the north. All these together make the aquifer a heterogeneous system.

A sandy clay/clay layer forms the base of the alluvial succession and is thought to be the top most part of the Chad Formation sediments. Its thickness was not resolved from the results of the present study, but it was seen to be continuous within the study site.

The Alluvial Groundwater Regime

A confined-unconfined groundwater regime exists within the present site and is an inherent characteristic of the alluvial groundwater system under its present undisturbed state. The spatial coordinates and the relative areal extents of the confined-unconfined regions are associated with variations in the overlying clay unit thickness and changes in the hydrologic status of the basin.

Temporal and spatial variations of water levels at 20 piezometers indicate that the hydrologic regime within the study site may be divided into several subregimes. The transitional boundaries between two adjacent subregimes can be inferred from the contour maps of water level elevation. Conclusions drawn from such maps may at times however become ambiguous for various reasons:

1) sampling distribution in space and time affects contouring;
2) uncertainty in aquifer stratigraphy and well perforation depth precludes grouping wells according to the aquifers being tapped, confined or unconfined;
3) multiple sources of recharge and discharge can perturb the natural flow pattern;
4) barriers are not necessarily perfect impermeable boundaries; and
5) hydraulic properties vary in three dimensions.

Because the effects of some of the above factors have been identified, the groundwater regime at the study site are grouped into the following (figure 6.9):

I) areas of immediate groundwater level response to increases in river stage. These areas encompass up to approximately 1000m from the river on the north side, and up to 700m from the river on the south side;

II) an intermediate area of perennially high water level; this area is found in the central portion (both north and south) of the floodplain. The south central region also receives localised vertical infiltration;

III) area with an outward gradient toward the floodplain margin. This area is adjacent to area II north of the river, and has a perennial hydraulic gradient toward the northern margin of the floodplain; and

IV) areas with negligible water level changes near the southern margin of the floodplain. Water level changes at these locations are insignificant, compared to water level changes in other parts of the study site.

The hydraulic status of areas I to IV on three dates, each in the wet and dry season respectively, are shown in table 6.4. It is evident from this table that some of these areas existed under more than one regime during the one year observation period (1992/93). The evolution of the groundwater regime was a function of time, and accordingly, the aquifer storativity did not remain constant throughout. For instance areas IN and IIS changed from an unconfined to confined state in the wet season, and subsequently returned to the unconfined state in the dry season. These changes evolved dynamically and the system at all times was characterised by a mixture of confined-unconfined regimes. Consequently the aquifer storativity grades spatially and its magnitude progressively alters from one which is large to another of much more smaller size, and vice versa.

Table 6.4. The alluvial groundwater regimes in the wet and dry seasons (1992/93)

	DATE					
Area	**25.6.92 (Dry)**	**22.7.92 (Wet)**	**22.8.92 (Wet)**	**15.9.92 (Wet)**	**13.11.92 (Dry)**	**27.6.93 (Dry)**
IN	Unconfined	Unconfined	Confined	Confined	Unconfined	Unconfined
IS	Confined	Confined	Confined	Confined	Confined	Confined
IIN	Confined	Confined	Confined	Confined	Confined	Confined
IIS	Unconfined	Unconfined	Unconfined	Confined	Confined	Unconfined
III	Unconfined	Unconfined	Unconfined	Confined	Unconfined	Unconfined
IV	Unconfined	Unconfined	Unconfined	Unconfined	Unconfined	Unconfined

N - North S - South

The facts outlined above, and the discussions given throughout, show that the alluvial aquifer system exhibits both confined and unconfined features simultaneously and, hence, it is clearly inadequate to describe a system with such an identity with a constant storativity (confined or unconfined) model. If such a model is imposed upon it, it would give an unreliable prediction of the practical reality (Rushton and Wedderburn, 1971; Rushton and Tomlinson, 1975). Use of a model with confined storativity, for instance, would possibly represent the confined regions of the aquifer adequately but would be likely to cause serious errors in the predicted heads for the unconfined areas by over-estimating the rise and fall of the water levels. It would be impossible to satisfy the two conditions simultaneously while using a single parameter. Such impossibility is highlighted in a study by Reynolds (1987) who attempted to analyse propagation of a single wave through an aquifer that changed from unconfined to confined state. He experienced difficulties in matching simulated response curves to observed response curves with both correct magnitudes and correct time lags. By varying diffusivity, he could make the time lag of the wave or the efficiency factor agree, but not both at the same time. By matching the lag time rather than the efficiency factor to the observed response, his simulated diffusivity tended to correspond to a confined storage value. Thus it is necessary that the presence and periodic adjustments in the status of the confined-unconfined regimes are recognised and adequately represented in a model that would be designed to represent the Yobe River-aquifer system. In addition in the unconfined zone it is necessary to include a non-linear function which allows the transmissivity to change with changes in the aquifer water level. Areal variations in transmissivity have also to be included to allow for the observed aquifer heterogeneity.

A practicable and adequate representation of an aquifer which possesses both confined and unconfined features, such as the present, can either be made in an analogue model by adapting the "automatic switching" concept of Rushton and Wedderburn (1971), or in a numerical model by following the procedures presented by Rushton and Tomlinson (1975).

The results of a detailed study of geological and hydrological processes such as the present must always be specific to some extent to the area studied. Nevertheless, because of the observed similarity in the disposition of the alluvium within the present site and those along the valley downstream (Water Surveys,1994), we suspect that a detailed monitoring of the aquifer water levels is likely to show conditions similar to the one obtained here. It is essential therefore to delineate the extent and the physical characteristics of the confined-unconfined regions and the inter-relationship between them regionally in relation to the evidence available and physically realistic conjections. Once this is successful, a realistic and practicable regional groundwater model would be attained.

Boundary Conditions

The Yobe River Channel

The Yobe is an underfit river and was seen to interact with the adjacent aquifer. Its channel is a recharge or discharge boundary to the alluvial groundwater system depending on the particular time of the year. The channel is a recharge boundary in the wet season and discharges groundwater in the dry season.

Discharge of aquifer water into the river channel takes place in the dry season when the river stage is below the aquifer water level. A notable feature of the aquifer water level in the dry season (figure 6.9) is its gradient towards the channel bed. Most of the water exchange between an underfit river and the adjacent aquifer takes place through the channel bed (Sharp,1977). Flow is dependent upon the hydraulic head difference between the river and the aquifer, and is affected by the thickness and permeability of the channel bed. Thus this boundary can be defined as a head dependent leakage boundary for which the thickness, area, vertical hydraulic conductivity of the channel bed, and the head in the aquifer close to the channel and river channel itself are specified/determined.

An alternative representation of head dependent leakage boundary in a model has been examined by Rushton and Tomlinson (1979). The relation between leakage and head difference was represented as 1) linear; 2) non-linear; and 3) a combination of 1 and 2. The combined linear-nonlinear expression of leakage was found to give a better representation of the observed aquifer behaviour. Thus adaptation of this concept would seem appropriate in describing the Yobe River-alluvial aquifer interaction in a modelling scheme.

Lower aquifer boundary

A continuous Chad clay/sandy clay layer forms the base of the alluvial aquifer and separates it from the Chad Formation aquifers below. However, it is not known whether the observed geologic separation also constitutes a hydraulic discontinuity between the alluvial groundwater and groundwater in the deeper Chad Formation aquifers. We do not have information on the thickness and the hydraulic properties of the Chad clay to verify the exact nature of the connection or separation influenced by it.

Hydraulic connection between the alluvial and deeper Chad aquifers may exist as a result of vertical water leakage through this unit. The degree of interaction is dependent upon the effective permeability of the Chad clay and the vertical hydraulic gradient. Such flow may be substantial at points where due to facies changes the leakage coefficient of the unit is enhanced, or because of lenticular distribution the Chad clay is absent allowing the two aquifers to be connected directly to each other. In such a case flow can take place from the alluvial system into the Chad Formations or vice versa depending upon the direction of the vertical hydraulic gradient. The alluvial system would be hydrogeologically separated from the lower aquifers if the Chad clay is very thick and its permeability is very small such that the magnitude of the vertical leakage is negligible. Both of these two concepts are plausible in a sedimentary basin in which two aquifers are vertically juxtaposed and separated by an aquitard. Thus the base of the alluvial aquifer can either be a no-flow or specified flow boundary.

Lateral Aquifer Boundaries

Northern boundary: Water level maps indicate that lateral flow from the alluvium towards the adjacent northern upland persisted throughout the year. Equally drill logs of the alluvium and the aquifer underlying the contiguous upland have similar facies and suggested geologic continuity (up to 2km inland) between the two units. However, it is not known whether this continuity is localised and restricted to a limited area, or extended to a considerable distance inland over a great width. In the absence of adequate information on the extent of this continuity, we can consider this boundary as a specified flow or specified gradient boundary. Detailed examination of this boundary may indicate otherwise.

Southern boundary: The exact nature of the boundary condition that skirt the south margin of the flood plain is also not known. However, flow towards the adjacent upland is seen to be much more restricted than on the north margin. This boundary could similarly be a specified gradient or flow boundary but one with a reduced flow.

It could also be a no-flow boundary if the alluvial sand and gravel unit pinches out or becomes too clay-dominated such that the flow magnitude is negligible across this boundary into the adjacent aquifer.

Summary and recommendations

Variations in aquifer storativity have been recognised as an important factor in understanding the hydraulic behaviour of the alluvial groundwater system. A study of the hydrogeology of the system within the present site enables several confined-unconfined groundwater regions to be defined based on their relative response to river stage changes. The spatial distribution of these regions is partly associated with the variation in the upper clay unit thickness, which is in turn associated with the depositional patterns of the fluvial sediments. Comparison of changes in river stage and groundwater levels in these regions in the dry and wet seasons enables a conceptual flow model of the river-aquifer system to be developed. A multi-layer, spatially distributed model is proposed, in which transitions between confined and unconfined conditions can be realistically represented. A numerical model needs to be designed with these concepts in mind in order to simulate the Yobe River-aquifer system. Once proven, the model can serve as a tool for the development and management of the aquifer.

We present the following recommendations to provide the basis for the development of such a model:

1) Determine the distribution of hydrostratigraphic unit thicknesses. Our observations have shown that the distribution of the upper clay unit thickness is particularly significant. The particular disposition of this unit controls aquifer thickness and geometry, and partly determines the state of the groundwater regime. In view of this:

   a) determine the distribution of the thickness and properties of the upper clay;

   b) determine if an unsaturated zone exists beneath the clay; and

c) determine the depth of flood ponding along with a hydrograph of the elevation of its surface. This will be useful in determining the magnitude of the infiltration recharge.

2) On the basis of (1), install piezometers at representative sites and determine the aquifer water level changes in response to river flow and flooding. Such sites should range from locations where the surface clay is very thick, to points where it is less thick or absent. Locations with intermediate clay thickness should also be considered. Variations in piezometric level response can possibly enable separation of channel seepage input from infitration. Also piezometric level records together with the determined upper unit thickness would adequately define the groundwater regime,i.e, confined or unconfined, both spatially and temporally.

3) Obtain accurate and reliable values of aquifer hydraulic properties by conducting several carefully controlled pumping tests. These should include long term tests designed to measure horizontal and vertical hydraulic conductivity and storativity. In order to accomplish this objective, the pump test wells should be designed and constructed such that some are open to the entire thickness of the aquifer. In some cases, tests should be conducted on separate hydrostratigraphic units. Sufficient time-drawdown data should be recorded such that the early, middle, and late portions of the response of the aquifer are defined. Time-drawdown data from long term pump tests are helpful in interpreting delayed yield, leakage and boundaries for calibrating and verifying the mathematical model. Long term tests should be in the order of one week or more, depending on the pumping rate and other factors such as boundaries and hydrostratigraphy as defined by the drilling data and geophysical surveys.

4) Delineate aquifer boundaries for model input during the field investigations. This can be achieved in the course of carrying out steps one to three above.

CHAPTER SEVEN

CONCLUSIONS AND RECOMMENDATIONS

7.1 Study conclusions

Conclusions about detailed aspects of the studies undertaken are given in the appropriate sections of the text. The following is a summary of the main conclusions drawn from each section:

1) it is revealed that the middle section of the Hadejia-Jama'are-Yobe River basin within the study site comprises alluvial deposits of variable composition; the stratigraphical sequence contains a sand and gravel layer sandwiched between two clay-rich units. At any particular location many, but not necessarily all, of the strata in the sequence are present, and both the vertical and lateral extents of the various strata are variable. The nature and disposition of these units conformed to a classical fluvial depositional pattern in a low energy (meandering) environment and were adequately described by a standard fluvial depositional model;

2) the floodplain surface is found to be covered at most locations by a continuous clay layer of more than 1.0m thick. The presence of this layer partly influences the alluvial aquifer hydraulic head distribution and the manner in which the aquifer responds to seasonal flood and recession. Where this unit is less than 1m thick, it allows localised vertical infiltration of flood water into the underlying aquifer;

3) the Yobe River is shown to be in hydraulic continuity with the adjacent alluvial aquifer and variations in the aquifer storativity have been recognised as an important factor in understanding the hydraulic behaviour of the Yobe River-aquifer system. A confined-unconfined groundwater regime exists within the present site and is an inherent characteristic of the alluvial groundwater system under its present undisturbed state. Most parts of the aquifer north of the river were confined throughout the year and were responsive to all changes in the river stage. The south sector on the other hand exhibited features akin to an unconfined aquifer, and

significant groundwater level increases there took place only during sustained high river stage;

4) the recognition of the differing response from the two sections of the aquifer to river stage changes enables a conceptual model of the system to be developed. A multi-layer, spatially distributed model is proposed, in which transitions between confined and unconfined conditions can be realistically represented. This model would serve as the framework on which to hang details, and can assist in obtaining pertinent data for the development of a numerical model of the river-aquifer system; and

5) the combination of geophysical technique, drilling and borehole water level measurements, and a conceptual model depicting the river-aquifer system, has given a better understanding of the river-aquifer interaction. It is for this reason that the research techniques employed in this study can be adapted for investigating the remainder of the basin downstream of the present study site.

7.2 Implications for small scale irrigation

The observed hydrogeological conditions show that the alluvial groundwater flow pattern represents a continuum of constantly changing conditions and adjustments. Water exchange between the river and aquifer takes place throughout the year and the aquifer at all times is characterised by a mixture of confined-unconfined regimes.

Following the preliminary findings of the present study, terms of reference were drawn up for NEAZDP by the present author, and hydrogeological investigations were carried out by Water Surveys (1994) for the areas downstream (between Gashua and Geidam) of the present study site. The results of the investigation show that the lower parts of the Yobe basin display a sedimentary sequence similar to that obtained in the present study; and most significant is that the aquifer is covered by a continuous clay layer at nearly all the locations considered. This revelation encourages one to speculate on the nature of the river-alluvial aquifer interaction and the hydraulic state of the aquifer downstream. The aquifer exists in a dynamic state and exhibits both confined and unconfined characteristics simultaneously. The recharge to the aquifer occurs through channel seepage and localised vertical infiltration, and its magnitude probably reduces with distance downstream (because of reducing river stage downstream). Sustainable development and utilisation of water resources in the lower part of the Yobe basin will intimately depend upon a prudent exploitation of the existing river-aquifer relationship.

The river-aquifer interaction observed will be disturbed when subject to groundwater abstractions. Generally alluvial aquifer water levels will be lowered when pumped for irrigation; and when this occurs in the vicinity of a river, the amount of groundwater returning to the river as baseflow is consequently reduced. Likewise lowering of the alluvial groundwater level also has the effect of increasing recharge to the aquifer by causing induced infiltration of river water. If such induced infiltration is to be feasible in the lower Yobe basin, it will form a major component of the groundwater development scheme for small scale irrigation. It would mean lowering the aquifer water level by pumping water for irrigation during the dry season and allowing it to recover during the wet season.

However, the feasibility of such a scheme in the basin will depend upon the availability of river water and the behaviour of the river-aquifer system when stressed. If the dewatered parts of the aquifer can recover fully during the flood season, the river-aquifer system will be self-regulating, and alluvial groundwater can safely be exploited and utilised for irrigation. Conversely, if the aquifer water levels were not able to recover and the residual drawdown continued through the wet season, it would imply that the system cannot bear the pumping stress imposed upon it and this could lead to undesirable cumulative effects in the following dry season and subsequent depletion of the entire system.

Currently the Yobe River flows only for a maximum period of 3 to 4 months, July to October, and remains effluent/dry for more than six months. Groundwater pumping for irrigation on the other hand takes place between November and February of each year. Thus the two processes are not in phase with one another and under the present condition manipulation of the river-aquifer system will depend entirely on the availability of annual river flow.

The period between cessation of river flow and commencement of pumping could be bridged by prolonging the length of river flow by releasing water from upstream impoundments. Such a scheme would maintain surface water flow and farmers could pump water from the aquifer, most of which is likely to be compensated for by induced infiltration.

In the absence of sufficient information on the foregoing, it is the author's view that, at this preliminary stage of exploring the river-aquifer relationship, the concept of large scale groundwater abstractions for irrigation should be postulated only as a long term aim. In the meantime pilot schemes need to be established at selected sites along the river, developing intensive irrigation deliberately aiming to lower the aquifer water level. The stressed river-aquifer system should then be monitored to see how it responds to the annual river flow. If the response is positive as described earlier, irrigation can then be expanded to other areas. If, however, the stressed aquifer could not recover completely the alternative scheme of releasing stored water should then be considered.

In considering this second scheme,

1) the rights of farmers who currently depend upon flood water for the cultivation of rice and recession crops must be respected. A scheme that prolongs river flow but completely takes away flooding "rights" will raise more controversy and be economically less viable to the downstream farmers. This is because flood water rice

cultivation is culturally entrenched into the society and the annual income currently obtained from flood rice cultivation far exceeds that from irrigated agriculture; and

2) the river stage and dam water release operation rules necessary for ensuring the viability of such a scheme must be scientifically determined and capable of being practically implemented. This will require the development of a comprehensive basinwide water management scheme.

7.3 Towards developing a basinwide water management scheme

The following aspects are important:

1) a new, integrated approach to conjunctive use of the basin's water resources, both surface and subsurface. Existing water resource development plans are generally concerned with either surface water or groundwater, and seldom give sufficient attention to conjunctive development and management. Both prudence in public policy and fiscal realities demand a unified management of the basin's water resources. This unified approach to water management would

a) provide the framework for anticipatory action for protecting the quality and quantity of surface and groundwater;

b) establish an institutional basis for integrating and targeting diverse programs; and

c) provide a means for judging effectiveness of programs for the sake of accountability.

2) If acceptance and implementation of a comprehensive water resource management program is to be achieved, the local people involved must understand the potential problem, believe in the need for action and appreciate that the benefits to be achieved by a management program outweigh the loss of individual choice.

In the Yobe basin enormous difficulties exist in establishing (2). For instance, water users in the upstream States are in a much more favourable position with respect to surface water resources and, thus, may be reluctant to be restricted in any way for the benefit of those downstream. Thus without full understanding and willingness, implementation of any management scheme would be politically and practically difficult and controversial. It is important that the formulation of a basinwide water management scheme incorporates the political, social, and economic framework existing in the region. The development and practical implementation of such a scheme would require the articulation of these factors, the development of management tools to allocate the total water resource fairly among the users, and initiation and carrying forward of actions that would ensure the viability of the scheme.

In Nigeria, the only authority that has both the resources and power to ensure this happens is the Federal Government and, therefore, its full participation will be required to establish the necessary local/state/federal partnership which would be involved in the management of the basin's water resources. Creation of such a partnership would require extensive discussions with States and local agencies involved; and the concurrence of most, if not all, would be necessary for any such management scheme to be politically and practically viable.

7.4 Further research needs

In the absence of sufficient data, a conceptual model was developed to describe the Yobe River-alluvial aquifer interaction; and without having quantitative relationships, qualitative propositions were put forward regarding the beneficial development of the Yobe basin water resource. The reason for so doing is not only to highlight the significance of the river-aquifer relationship hitherto disclosed, but also an attempt to bring into perspective the uncertainties, difficulties and challenges the development and management of this limited, invaluable, natural asset of the region presents.

In view of this the following are recommended for in-depth study and development:

Field studies

1) The present study provided only a localised, albeit detailed, picture of the Yobe basin alluvium in terms of its geology and hydrogeology. There is, therefore, the need to carry out extensive field work in the areas downstream of the present study site to obtain a clearer and more complete picture of the basin alluvium. These studies should aim, using the techniques used in the present study

a) to define the alluvium stratigraphy regionally;

b) to determine aquifer hydraulic parameters namely transmissivity and storativity; and

c) to monitor aquifer and river water levels.

In performing the foregoing it is strongly recommended that all the studies should be carried out at the most representative sites, and not randomly. This is very important because optimising the amount of field work involved would make the operation less costly and more effective. The initial step would, therefore, be to determine the optimal test sites and monitoring networks using the available data, such as those obtained from the present work and previous studies. With these sites established, the field investigations can be carried out effectively as part of a routine technical study while adopting the combined approach used in the present study. Geostatistical techniques such as kriging is a useful tool in making such designs.

2) A potentially promising approach for selecting optimal sites for field investigation involves the identification of locations where localised vertical flood water infiltration may be occuring. Such a procedure has been suggested recently by Marinof-Petkoff (1994). He undertook aerial photograph interpretations, for a section of the floodplain downstream of the present site, as part of his Msc project and postulated that areas

identified as sand bars and terraces are possible sites for vertical flood water infiltration. This hypothesis has not been tested, and, thus, its verification in the field will be useful.

3) Field studies which are equally important and need to occur in tandem with that of the alluvium are those examining the nature, both laterally and vertically, of the relationship between the alluvial aquifer and the underlying/contiguous Chad Formation aquifers. Studies addressing this relationship are recommended.

Modelling and model application

There is the need to design a comprehensive, distributed parameter numerical model of the combined river-aquifer system. The development of the model should commence soon, using the conceptual model developed in the present study as a basis, and progress in tandem with the development program. Such a model will make a desirable management tool for predicting responses and optimising the development and use of the total water resources.

In the meantime existing models need to be adopted and modified, and applied to simulate the system behaviour under different development and management scenarios. The same model can also be used to establish the necessary river hydrograph which can ensure a complete annual groundwater recharge and other requirements such flooding.

Irrigation development

The background upon which the present study was commenced is the pressing need to have a base upon which small scale irrigation in the Yobe valley, in particular in the NEAZDP area, can be planned and undertaken without impacting undesirable effects on the system. The findings of the present study show that the water resource of the valley is in a delicate dynamic balance. Because of this extensive irrigation development in the NEAZDP area should be planned with care and taken as a long term aim. In the meantime selective irrigation schemes can be established and intensive irrigation carried out with careful monitoring of the river-aquifer system response to the pumping stress and annual river flow. It is recommended that four sites, one each near Gashua and Geidam and the other two downstream of Gashua, be established for pilot irrigation schemes. Each site should contain wells which are purposely set aside for water level monitoring. The information obtained from these pilot schemes would serve as the baseline from which appropriate development strategies necessary for the prudent utilisation of the system can be designed.

One of the difficulties encountered throughout the present study is the absence of adequate data base the study could have relied upon. All the water institutions and agencies visited by the author during field work for information gathering lack comprehensive classification and standards for borehole logging and aquifer monitoring. Quality of groundwater data is commonly poor, systematic data collection efforts are absent, data interpretation is limited and files of data are virtually inaccessible. NEAZDP, to be comprehensive in its groundwater development and irrigation schemes, should place adequate effort in collecting and keeping records and information of borehole locations and logging, and other informations such as groundwater levels.

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APPENDIX A

PIEZOMETER INSTALLATION DETAILS

Appendix A. Levels of piezometer installation in relation to ground surfaces

Piezometer	Distance from the river (m)	Ground surface elevation (M-a.s.l.)	Piezometer tip elevation (M-a.s.l.)
P_1	50	333.91	319.41
P_2	130	333.73	321.53
P_3	400	332.95	319.95
P_4	800	333.09	320.09
P_5	1140	333.90	319.90
P_6	2100	333.96	319.46
P_7	2700	333.40	321.90
P_8	2900	334.37	321.37
P_9	50	334.00	318.00
A	140	334.30	321.30
B	400	334.71	320.21
C	650	333.57	319.07
D	900	333.93	319.33
E	1140	334.34	321.34
F	1400	333.70	320.70
G	1900	333.72	320.72
H	3100	334.24	319.74
I	2250	333.77	322.77
J	250	333.89	322.89
K	250	333.72	320.73

APPENDIX B

WATER LEVEL DATA

Appendix B. Piezometer data: Yobe River floodplain 1992/93.
Water level read referenced to each datum (m)

Piezometer	Ground surface elevation (M-a.s.l)	1992												
		13/5 1	28/5 15	9/6 27	25/6 43	4/7 52	9/7 57	22/7 70	22/8 101	15/9 125	29/10 169	13/11 184	30/11 201	11/12 212
P_1	333.91	3.62	3.69	3.71	3.78	3.66	3.57	3.37	1.07	-0.12	0.89		2.12	1.78
P_2	333.73	3.34	3.40	3.43	3.50	3.43	3.38	3.17	0.61	-0.43	0.60	1.09	1.43	1.63
P_3	332.95	2.53	2.56	2.58	2.62	2.63	2.64	2.45		-1.15	-0.23	0.32	0.64	0.78
P_4	333.09	2.46	2.50	2.53	2.56	2.58	2.58	2.59	0.80	-0.43	-0.60	-0.22	0.23	0.14
P_5	333.90	2.26	2.31	2.39	2.47	2.52	2.54	2.59	2.35	0.45	-0.10	0.22	0.48	0.61
P_6	333.46	2.60	2.79	2.77	2.80	2.80	2.79	2.77	2.70	-0.35	-0.72	-0.65	0.29	0.92
P_7	333.40	5.80	5.90	5.91	6.02	6.07	6.00	5.95	5.66	4.85	4.52	4.61	4.73	4.75
P_8	334.37	6.80	6.80	7.00	6.80	6.80	6.08	7.01	6.67	6.09	5.75	5.82	5.88	5.89
P_9	334.00	3.89	3.92	3.94	3.98	3.70	3.49	3.06	1.00	-0.03	1.24	2.29	2.84	2.84
A	334.30	3.75	3.80	3.82	3.86	3.76	3.60	3.18	0.74	-0.15	1.05	1.87	2.34	2.51
B	334.71	3.98	4.01	4.04	4.07	3.73	3.88	3.56	1.75	0.29	1.49	2.19	2.59	2.75
C	333.57	2.76	2.80	2.83	2.85	2.79	2.70	2.42	0.50	-0.67	0.34	1.01	1.39	1.55
D	333.93	3.27	3.31	3.34	3.37	3.35	3.29	3.10	1.14	-0.08	0.94	1.49	1.87	2.03
E	334.34	3.39	3.40	3.25	3.46	3.51	3.49	3.41	2.00	0.36	1.09	1.69	2.05	2.18
F	333.70	2.83	2.81	2.86	2.94	2.95	2.97	2.94	1.83	-0.05	0.68	1.25	1.62	1.75
G	333.72	2.74	2.96	2.98	3.04	3.08	3.09	3.07	1.17	0.34	0.84		1.78	1.88
H	334.24	4.30	4.31	4.42	4.47	4.49	4.50	4.51	2.96	1.73	2.26	2.76	3.07	3.18
I	333.77	2.34	2.36	2.47	2.49	2.49	2.53	2.61	2.30	0.22	-0.23	-0.26	0.23	1.32
J	333.89	3.05	3.10	3.12	3.14	3.05	2.92	2.64	-0.18	-0.97	-0.66	1.42	1.78	1.91
K	333.72	3.04	3.42	3.37	3.46	3.42	3.34	3.14	1.87	0.21	1.077	1.84	2.19	2.32

13/5 - date (13/5/92)   70 - cumulative days from 13/5/92
(-0.12) - water level above ground surface

Appendix B. Continued

Piezometer	Ground surface elevation (M-a.s.l)	1993												
		1/1 233	15/1 247	29/1 261	13/2 276	26/2 289	15/3 306	25/3 316	19/4 341	29/4 351	18/5 370	29/5 381	1/6 384	27/6 410
P_1	333.91	2.84	2.83	2.82	2.84	2.78	2.84	2.84			2.68	2.59	2.61	2.64
P_2	333.73	2.07	2.29	2.42	2.59	2.65	2.83	2.88						
P_3	332.95	1.14	1.35	1.47	1.65	1.70	1.89	1.97	2.12	2.17	2.28	2.27	2.29	2.32
P_4	333.09	0.34	0.42	0.45	0.54	0.55	0.68	0.75	0.86	0.92	1.65	1.08	1.13	1.25
P_5	333.90	0.84	0.99	1.10	1.25	1.30	1.45	1.52	1.66	1.71	1.81	1.84	1.87	2.03
P_6	333.46	0.98	1.02	0.27	0.34	0.47	0.51	0.22	0.27	0.40	0.48	0.21	0.16	0.31
P_7	333.40	4.88	4.92	4.98	5.06	5.19	5.23	5.29	5.39	5.42	5.48	5.55	5.56	5.70
P_8	334.37	5.99	6.02	6.05	6.12	6.14	6.24	6.28	6.42	6.48	6.39	6.55	6.58	6.71
P_9	334.00	3.12	3.25	3.34	3.45	3.52	3.59	3.65	3.72	3.76	3.86	3.84	3.85	3.46
A	334.30	2.81	2.97	3.07	3.22	3.27	3.38	3.44	3.55	3.56	3.65	3.59	3.60	3.60
B	334.71	3.03	3.17	3.27	2.88	3.46	3.58	3.64	3.75	3.76	3.66	3.67		
C	333.58	1.83	1.96	2.07	2.19	2.24	2.36	2.42	2.52	2.52				
D	333.93	2.34	2.47	2.98	2.69	2.73		2.90	2.94	3.05	3.09	3.09	3.13	3.09
E	334.34	2.39	2.54	2.60	2.73	2.78	2.85	2.93	3.04	3.10	3.05	3.17	3.18	3.28
F	333.70	1.98	2.08	2.16	2.27	2.29	2.40	2.44	2.58	2.60	3.11	2.64	2.69	2.72
G	333.72	2.13	2.23	2.28	2.43	2.38	2.49	2.55	2.68	2.71	2.78	2.79	2.80	2.84
H	334.24	3.39	3.43	3.53	3.72	3.75	3.86	3.92	4.04	4.07	4.11	4.17	4.23	4.30
I	333.77	1.37	1.54	0.04	0.19	0.23	0.42	0.21	0.46	0.36	0.43	0.47	0.56	0.30
J	333.89	2.16	2.29	2.40	2.53	2.59	2.72	2.76	2.86	2.88	2.94	2.93	2.98	2.76
K	333.72	2.54	2.65	2.74	2.88	2.85	2.99	3.03	3.07	3.20	3.45	3.24	3.22	3.16

15/1 - date (15/1/93)   233 - cumulative days from 13/5/92
(-0.12) - water level above ground surface

Appendix B. Continued

Piezometer	Ground surface elevation (M-a.s.l)	1992												
		13/5 1	28/5 15	9/6 27	25/6 43	4/7 52	9/7 57	22/7 70	22/8 101	15/9 125	29/10 169	13/11 184	30/11 201	11/12 212
P_1	333.91	330.29	330.22	330.20	330.13	330.25	330.34	330.54	332.84	334.03	333.02		331.79	332.13
P_2	333.73	330.39	330.33	330.30	330.23	330.30	330.35	330.56	333.12	334.16	333.13	332.64	332.30	332.10
P_3	332.95	330.42	330.39	330.37	330.33	330.32	330.31	330.50		334.10	333.18	332.63	332.31	332.17
P_4	333.09	330.63	330.59	330.56	330.53	330.51	330.51	330.50	332.29	333.52	333.69	333.31	332.86	332.95
P_5	333.90	331.64	331.59	331.51	331.43	331.38	331.36	331.31	331.55	333.45	334.00	333.68	333.42	333.29
P_6	333.46	330.86	330.67	330.69	330.66	330.66	330.67	330.69	330.76	333.81	334.18	334.11	333.17	332.54
P_7	333.40	327.60	327.50	327.49	327.38	327.33	327.40	327.45	327.74	328.55	328.88	327.58	328.67	328.65
P_8	334.37	327.57	327.57	327.37	327.57	327.57	328.29	327.36	327.70	328.28	328.62	329.76	328.49	328.48
P_9	334.00	330.11	330.08	330.06	330.02	330.30	330.51	330.94	333.00	334.03	332.76	331.71	331.16	331.16
A	334.30	330.55	330.50	330.48	330.44	330.54	330.70	331.12	333.56	334.45	333.25	332.43	331.96	331.79
B	334.71	330.73	330.70	330.67	330.64	330.98	330.83	331.15	332.96	334.42	333.22	332.52	332.12	331.96
C	333.57	330.81	330.77	330.74	330.72	330.78	330.87	331.15	333.07	334.24	333.23	332.56	332.18	332.02
D	333.93	330.66	330.62	330.59	330.56	330.58	330.64	330.83	332.79	334.01	332.99	332.44	332.06	331.90
E	334.34	330.95	330.94	331.09	330.88	330.83	330.85	330.93	332.34	333.98	333.25	332.65	332.29	332.16
F	333.70	330.87	330.89	330.84	330.76	330.75	330.73	330.76	331.87	333.75	333.02	332.45	332.08	331.95
G	333.72	330.98	330.76	330.74	330.68	330.64	330.63	330.65	332.55	333.38	332.88		331.94	331.84
H	334.24	329.94	329.93	329.82	329.77	329.75	329.74	329.73	331.28	332.51	331.98	331.48	331.17	331.06
I	333.77	331.43	331.41	331.30	331.28	331.28	331.24	331.16	331.47	333.55	334.00	334.03	333.54	332.45
J	333.89	330.84	330.79	330.77	330.75	330.84	330.97	331.25	334.07	334.86	334.55	332.47	332.11	331.98
K	333.72	330.68	330.30	330.35	330.26	330.30	330.38	330.58	331.85	333.51	332.64	331.88	331.53	331.40

13/5 - date (13/5/92) 70 - cumulative days from 13/5/92

Appendix B. Continued

Piezometer	Ground surface elevation (M-a.s.l)	1993												
		1/1 233	15/1 247	29/1 261	13/2 276	26/2 289	15/3 306	25/3 316	19/4 341	29/4 351	18/5 370	29/5 381	1/6 384	27/6 410
P_1	333.91	331.07	331.08	331.09	331.07	331.13	331.07	331.07			331.23	331.32	331.30	331.27
P_2	333.73	331.66	331.44	331.31	331.14	331.08	330.90	330.85						
P_3	332.95	331.81	331.61	331.48	331.30	331.25	331.06	330.98	330.83	330.78	330.67	330.68	330.66	330.63
P_4	333.09	332.75	332.67	332.64	332.55	332.54	332.41	332.34	332.23	332.17	331.44	332.01	331.96	331.84
P_5	333.90	333.06	332.91	333.90	332.65	332.60	332.45	332.38	332.24	332.19	332.09	332.06	332.03	331.87
P_6	333.46	332.48	332.44	333.19	333.12	332.99	332.95	333.24	333.19	333.06	332.98	333.25	333.30	333.15
P_7	333.40	328.52	328.48	328.42	328.34	328.21	328.17	328.11	328.01	327.98	327.92	327.85	327.84	327.70
P_8	334.37	328.38	328.35	328.32	328.25	328.23	328.13	328.09	327.95	327.89	327.98	327.82	327.79	327.66
P_9	334.00	330.88	330.75	330.66	330.55	330.48	330.41	330.35	330.28	330.24	330.14	330.16	330.15	330.54
A	334.30	331.49	331.33	331.23	331.08	331.03	330.92	330.86	330.75	330.74	330.65	330.71	330.70	330.70
B	334.71	331.68	331.54	331.44	331.83	331.25	331.13	331.07	330.96	330.95	331.05	331.04		
C	333.58	331.74	331.61	331.50	331.38	331.33	331.21	331.15	331.05	331.05				
D	333.93	331.59	331.46	330.95	331.24	331.20		331.03	330.99	330.88	330.84	330.84	330.80	330.84
E	334.34	331.95	331.80	331.74	331.61	331.56	331.49	331.41	331.30	331.24	331.29	331.17	331.16	331.06
F	333.70	331.72	331.62	331.54	331.43	331.41	331.30	331.26	331.12	331.10	330.59	331.06	331.01	330.98
G	333.72	331.59	331.49	331.44	331.29	331.34	331.23	331.17	331.04	331.01	330.94	330.93	330.92	330.88
H	334.24	330.85	330.81	330.71	330.52	330.49	330.38	330.32	330.21	330.17	330.13	330.07	330.01	329.94
I	333.77	332.40	332.21	333.73	333.58	333.54	333.35	333.56	333.31	333.41	333.34	333.30	333.21	333.47
J	333.89	331.73	331.60	331.49	331.36	331.30	331.17	331.13	331.03	331.01	330.95	330.96	330.91	331.13
K	333.72	331.18	331.07	330.98	330.84	330.87	330.73	330.69	330.65	330.52	330.27	330.48	330.50	330.56

15/1 - date (15/1/93) 233 - cumulative days from 13/5/92

APPENDIX C

PUMPING TEST DATA

Appendix C. **Pumping test data at the study site**

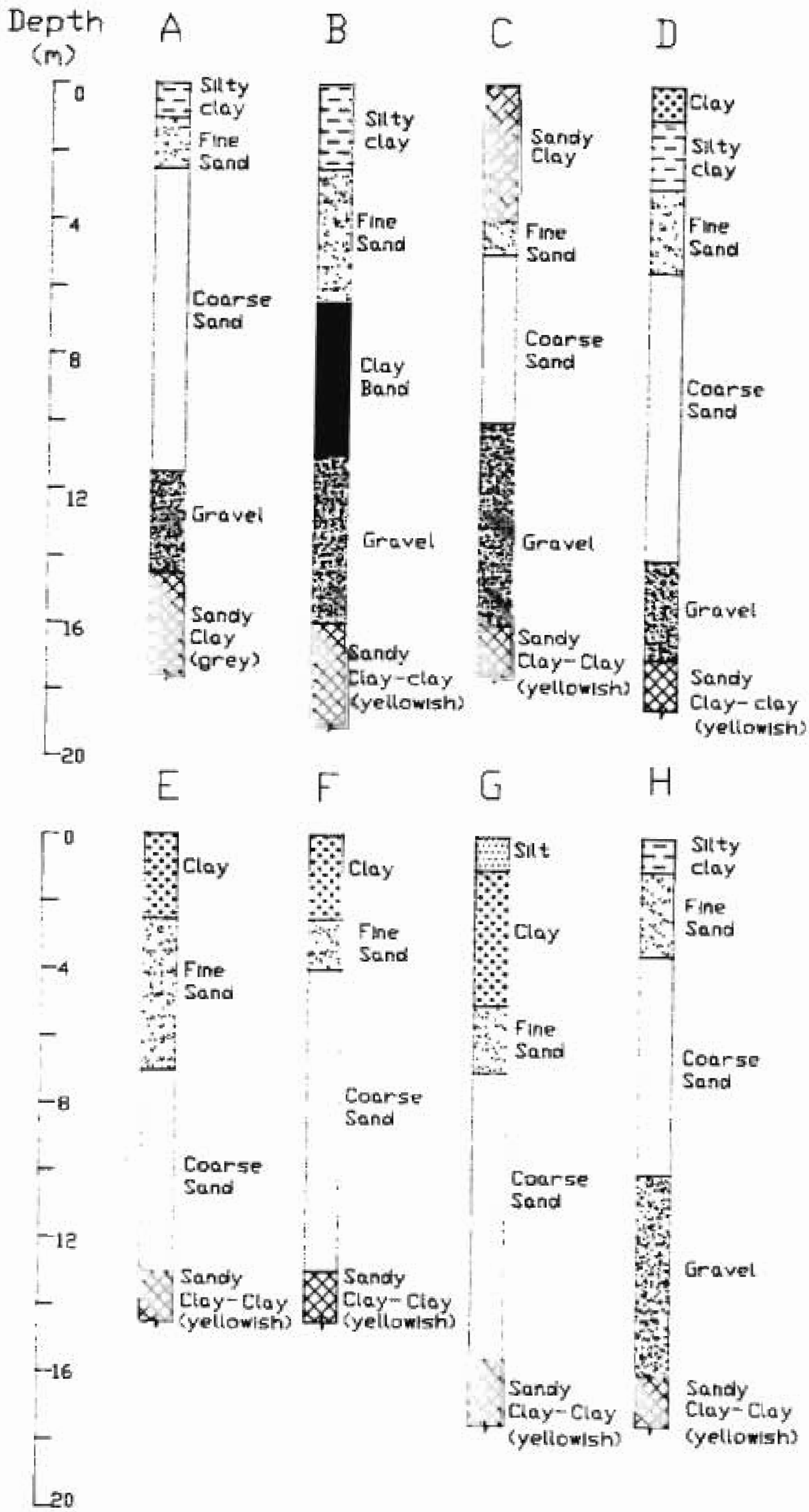
Time (minutes)	Observation well Drawdown (cm)	Observation well Recovery (cm)	Pumped well Drawdown (cm)	Pumped well Recovery (cm)
1	33.5	22.5		
2	37.5	27.5		
3	41.5	30.0		
4	41.5	31.5		
5	34.5	32.5		
6	36.5	33.0		
7	36.5	34.5		
8	36.5	35.5		
9	36.5	36.0		
10	37.0	36.5		185.0
11			181.0	
20	40.5	40.0	183.0	190.0
30	44.5	42.0	191.0	192.0
40	44.5	43.5	192.0	193.0
50	46.5	44.5	191.5	194.0
60	47.5	45.0	194.0	195.0
80	48.0	46.0	194.0	196.0
100	49.5		196.0	
190		49.0		199.0
200	52.5		204.0	

(Diameter of pumped well = 100mm; pumping rate = 142m^3/day; observation well distance = 4.30m)

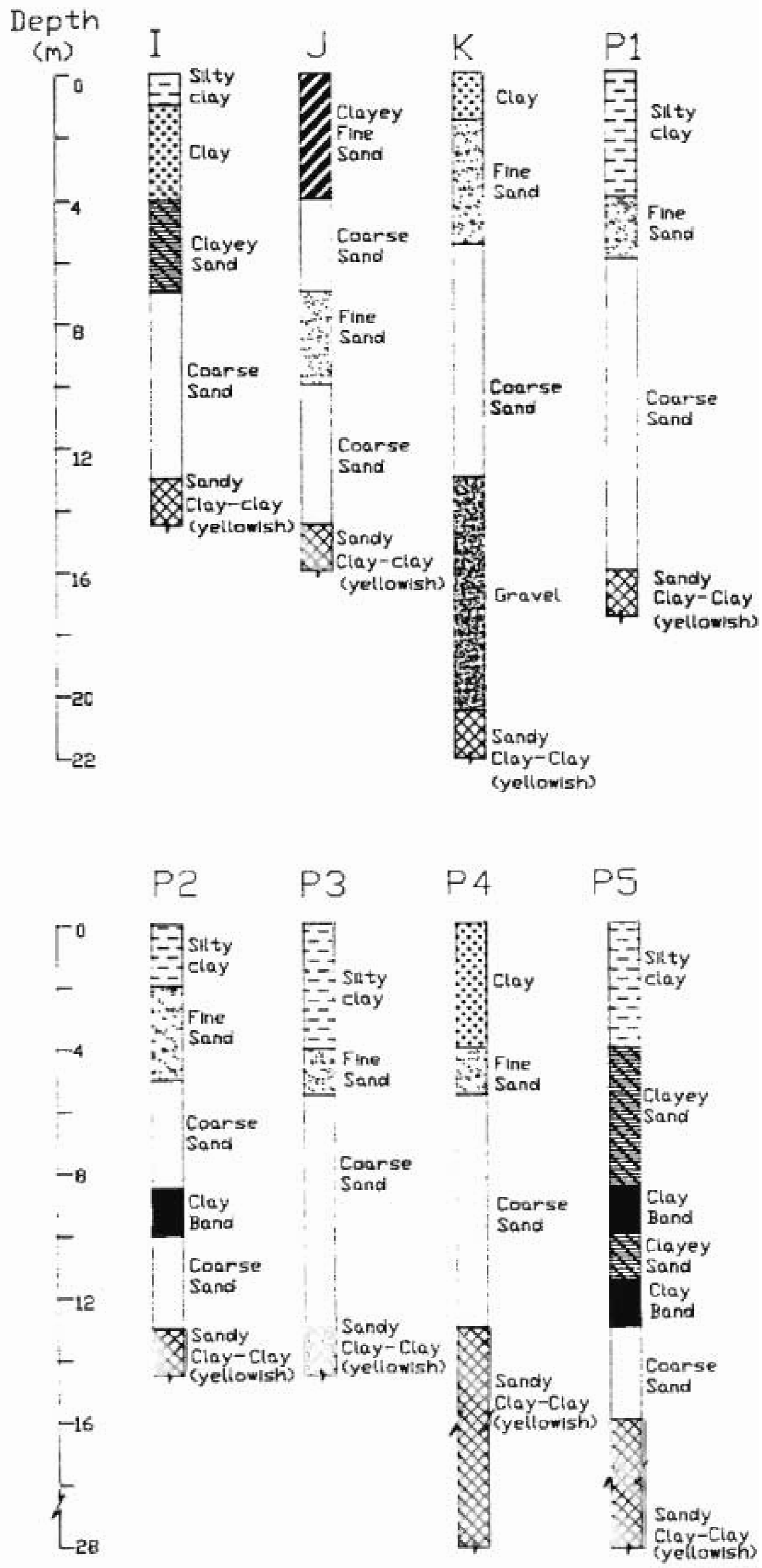
APPENDIX D

BOREHOLE LOGS

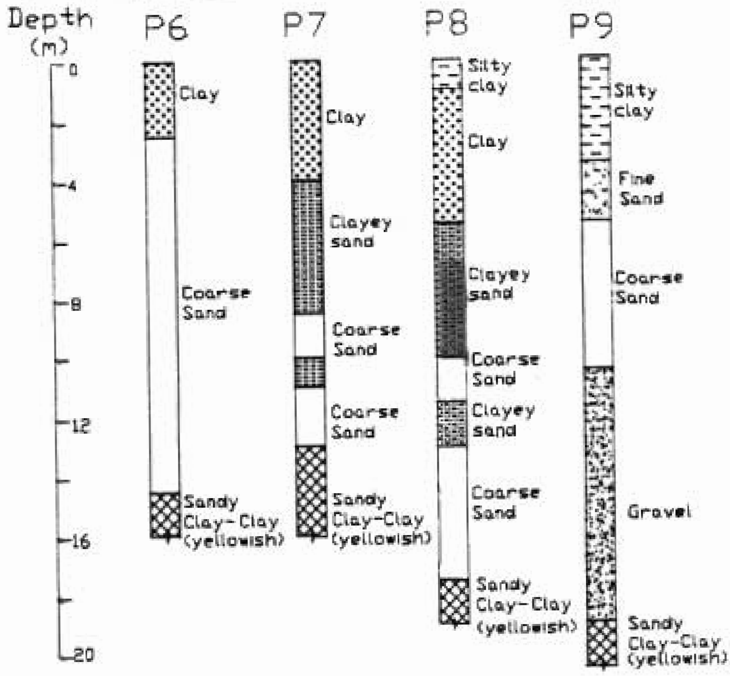
Appendix D. Borehole logs

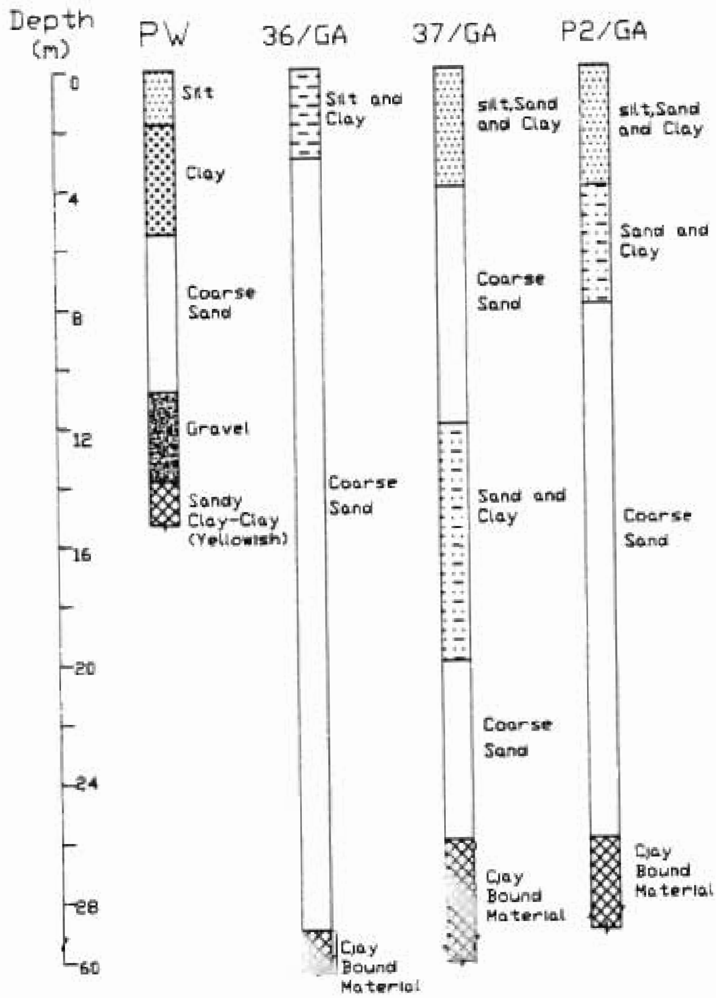


Appendix D. Continued



Appendix D Continued





APPENDIX E

VERTICAL ELECTRICAL SOUNDING (VES) DATA

Appendix E. Layer thicknesses and resistivities from automatic interpretation procedure

VES Station	Surface layer		Clay unit		Sand and gravel		Under-lying clay	Inter-pretation closure
	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	R (Ohm-M)	error %
1			0.35	34.20	21.64	250.00	18.00	8.78
2†								
3	0.27	47.83	1.62	12.65	10.07	103.10	33.46	8.11
4			0.43	76.80	17.77	313.60	30.19	8.74
5	0.66	11.00	1.58	4.84	9.32	268.80	35.12	8.00
6	0.34	72.18	2.68	8.90	12.88	293.00	16.90	5.18
7	0.34	18.00	1.66	8.50	10.00	350.00	25.00	5.12
8	0.73	22.91	2.58	11.30	9.94	236.40	52.10	7.72
9*			2.50	7.00	10.00	300.00	11.50	8.40
10*			1.87	12.59	18.74	254.40	9.47	7.89
11†								
12			3.14	13.55	8.43	275.20	1.77	7.83
13	0.33	11.86	2.21	6.95	10.91	236.10	14.75	5.87
14			0.26	16.00	13.74	180.00	34.50	7.49
15	0.32	25.50	2.18	11.00	11.50	175.00	18.00	6.58
16†								
17			0.70	29.23	20.30	240.00	78.00	9.78
18			1.14	10.33	16.36	338.40	6.74	9.44
19*			1.25	20.00	19.00	131.00	37.00	8.96
20*			0.67	17.32	19.80	188.80	15.82	9.16
21	0.40	52.00	4.10	14.00	10.50	220.00	25.00	8.98
22	0.28	29.52	1.65	16.83	12.53	305.00	9.39	9.38
23†								
24			1.31	17.47	9.11	280.50	11.45	8.51
25	0.23	33.71	3.04	13.50	8.06	192.80	7.00	7.45
26			0.75	15.50	15.70	120.00	38.00	5.78
27*			1.45	10.00	18.55	100.00	30.00	9.04
28*			9.00	64.00	19.00	100.00	3.00	9.92
29*			1.43	9.64	13.97	256.40	21.10	9.33
30			0.51	16.35	12.96	318.00	5.62	7.55
31*			2.50	12.50	10.50	190.00	44.00	7.58
32*			4.50	25.00	10.50	200.00	24.00	8.52
33*			1.80	9.00	10.70	200.00	10.00	9.83
34†								
35			0.40	56.00	10.00	342.00	39.90	9.44

H = Layer Thickness; R = Layer Resistivity; * indicates computations constrained by lithologic data from drill logs; and † VES Stations with spurious data.

Appendix E. (cont'd)

VES Station	Surface layer		Clay unit		Sand and gravel		Underlying clay	Interpretation closure error
	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	R (Ohm-M)	%
36			0.36	17.24	11.14	410.00	13.00	9.51
37			0.27	19.25	10.93	414.20	11.92	8.16
38†								
39			5.50	23.50	11.00	230.00	9.00	9.63
40			1.70	10.59	11.29	230.00	2.51	6.61
41			3.00	20.00	9.50	200.00	11.00	6.76
42			0.55	29.00	11.45	375.00	24.00	9.29
43			0.14	66.00	13.06	187.00	34.00	9.82
44			0.30	88.00	22.00	186.00	7.00	9.75
45	0.35	29.91	2.89	14.68	7.80	336.60	19.19	6.69
46			1.68	13.81	6.58	339.60	4.19	9.88
47	1.55	291.40	6.26	48.31	20.35	100.00	6.96	9.38
48			0.43	103.00	13.85	302.80	28.33	9.54
49			0.99	73.17	9.15	246.50	26.12	9.55
50			2.35	50.00	15.15	150.00	4.93	7.97
51	0.48	64.16	1.04	20.01	13.15	202.20	23.94	9.28
52	1.99	22.03	4.97	13.83	8.36	426.80	5.70	9.15
53	1.30	60.50	4.01	10.47	9.31	149.30	9.06	8.83
54*			0.86	37.86	10.54	310.10	3.00	8.77
55*			1.10	26.71	10.39	296.00	2.84	8.62
56			0.50	50.00	15.30	325.00	12.00	9.55
57			0.85	90.00	12.15	450.00	12.00	9.58
58			0.51	109.00	12.07	457.30	1.84	4.81
59			0.50	66.24	12.88	455.10	22.00	9.56
60	0.96	93.65	1.64	26.58	11.70	200.00	7.00	4.26
61	0.34	27.13	0.77	14.86	22.74	34.07	12.99	3.80
62			0.50	19.87	15.99	63.73	15.90	9.30
63			1.59	18.61	15.27	169.20	30.09	6.36
64*			1.00	12.00	9.00	350.00	14.00	8.18
65			3.50	80.00	14.00	215.00	11.00	9.14
66*			4.00	73.82	10.50	328.80	2.95	8.36
67			1.07	9.89	13.26	199.30	7.10	4.93
68	0.27	12.64	3.26	10.39	11.46	175.00	23.00	8.85
69			0.57	23.73	10.27	299.90	13.81	8.14
70	0.53	18.19	1.84	9.94	12.62	375.00	55.00	8.46

H = Layer Thickness; R = Layer Resistivity; * indicates computations constrained by lithologic data from drill logs; and † VES Stations with spurious data.

Appendix E. (cont'd)

VES Station	Surface layer		Clay unit		Sand and gravel		Underlying clay	Interpretation closure error
	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	H(m)	R (Ohm-M)	R (Ohm-M)	%
71			0.32	25.57	14.98	208.50	15.62	7.78
72			1.60	31.00	14.40	200.00	28.00	6.40
73			0.52	27.53	12.95	311.90	23.47	1.65
74			1.05	11.79	13.94	330.00	20.80	9.73
75	0.39	38.66	0.62	4.29	8.99	330.00	10.00	9.52
76			1.19	28.78	15.78	205.00	12.00	9.63
77†								
78			1.17	11.39	9.99	230.20	9.55	7.92
79†								
80			0.85	16.50	19.15	58.00	20.00	8.79
81			0.74	53.46	10.51	228.70	12.12	8.76
82			0.76	6.49	11.98	103.10	20.91	9.60
83			0.89	18.79	9.91	225.00	9.00	9.93
84			1.79	13.36	12.70	105.00	13.64	9.61
85			0.55	40.00	17.45	225.00	9.00	9.93
86*			5.00	40.00	11.00	135.00	35.00	7.41
87*	0.45	17.65	1.89	14.05	13.64	145.00	12.00	8.15
88			0.95	18.34	13.31	365.70	35.66	5.21
89			0.51	76.84	9.14	432.80	39.10	9.64
90	0.37	13.38	1.76	8.32	7.22	352.00	1.50	6.64
91			1.50	26.27	10.50	206.30	4.71	9.45
92			0.35	73.00	22.15	140.00	14.00	7.10
93†								
94			1.23	26.23	13.65	221.40	8.75	7.47
95†								
96*			1.49	13.00	12.00	225.00	5.00	6.98
97			0.52	55.38	11.27	164.60	23.44	8.86
98			0.25	35.60	11.55	368.00	10.27	9.06
99†								
100*			0.85	39.52	12.15	250.20	12.59	8.81
101	1.36	145.70	4.25	74.08	8.65	285.10	6.33	3.64
102†								

H = Layer Thickness; R = Layer Resistivity; * indicates computations constrained by lithologic data from drill logs; and † VES Stations with spurious data.

Appendix E2. Vertical Electric Sounding (VES) data: Yobe River Floodplain 1991/92

VES 1

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	25.4	1.35	24.3	12.95	21.00	16.98
1.0	21.0	1.32	19.84	15.66	12.06	13.86
2.0	16.6	1.06	15.501	10.80	11.96	11.38
4.0	10.30	0.77	9.55	7.90	7.35	7.63
8.0	5.24	0.54	4.89	4.30	3.92	4.11
16.0	2.88	0.13	2.70	2.22	2.07	2.15
32.0	0.69	0.05	0.63	0.72	0.68	0.70
64.0	0.12	0.00	0.12	0.11	0.10	0.11
128.0	0.03	0.00	0.03	0.02	0.02	0.02

VES 2

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	433.0	76.00	357.0	368.00	453.00	410.50
1.0	169.2	5.90	163.30	111.80	172.30	142.05
2.0	39.0	0.70	38.30	32.60	35.50	34.05
4.0	8.88	0.07	8.74	3.01	10.25	6.63
8.0	0.72	0.10	0.61	0.44	0.27	0.36
16.0	0.49	0.02	0.47	0.25	0.30	0.28
32.0	0.31	0.00	0.30	0.17	0.23	0.20
64.0	0.08	0.01	0.07	0.08	0.08	0.08
128.0	0.06	0.02	0.04	0.02	0.02	0.02

VES 3

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	14.52	0.06	14.80	10.24	6.60	8.42
1.0	2.50	0.28	2.19	3.50	1.01	2.26
2.0	2.58	0.16	2.46	1.95	1.23	1.59
4.0	1.84	0.13	1.70	1.43	1.16	1.30
8.0	1.20	0.09	1.07	0.93	0.77	0.85
16.0	0.79	0.05	0.71	0.60	0.51	0.56
32.0	0.49	0.01	0.48	0.27	0.28	0.28
64.0	0.14	0.02	0.12	0.10	0.10	0.10
128.0	0.06	0.00	0.05	0.03	0.02	0.03

VES 4

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	50.70	2.02	48.40	31.00	36.40	33.70
1.0	33.00	1.94	31.10	21.50	22.80	22.15
2.0	24.30	1.83	22.50	15.42	18.59	17.01
4.0	16.46	0.94	15.56	9.78	13.58	11.68
8.0	6.83	0.49	6.38	4.88	6.04	5.46
16.0	3.35	0.10	3.27	1.99	2.94	2.47
32.0	0.86	0.03	0.82	0.57	0.69	0.63
64.0	0.10	0.00	0.10	0.11	0.11	0.11
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 5

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	3.94	0.38	3.56	2.36	4.29	3.33
1.0	1.97	0.09	1.88	1.26	1.57	1.42
2.0	1.02	0.06	0.96	0.71	0.62	0.67
4.0	0.88	0.05	0.82	0.56	0.53	0.55
8.0	0.80	0.05	0.75	0.55	0.46	0.51
16.0	0.70	0.04	0.67	0.48	0.44	0.47
32.0	0.37	0.07	0.30	0.28	0.29	0.29
64.0	0.22	0.02	0.19	0.12	0.12	0.12
128.0	--	--	--	--	--	--

VES 6

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	15.59	0.58	14.97	13.72	10.92	12.32
1.0	2.63	0.15	2.51	2.63	2.19	2.41
2.0	1.41	0.07	1.35	1.01	0.84	0.93
4.0	0.98	0.08	0.89	0.64	0.69	0.67
8.0	0.98	0.07	0.91	0.69	0.66	0.68
16.0	0.68	k0.06	0.62	0.53	0.46	0.50
32.0	0.49	0.07	0.44	0.36	0.35	0.36
64.0	0.12	0.00	0.12	0.15	0.13	0.14
128.0	0.05	0.01	0.04	0.03	0.02	0.03

VES 7

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.20	0.24	5.96	4.90	3.84	4.37
1.0	2.31	0.13	2.19	2.12	1.36	1.74
2.0	1.60	0.08	1.52	1.12	0.95	1.04
4.0	1.50	0.08	1.38	0.87	0.95	0.91
8.0	1.38	0.09	1.29	0.86	0.91	0.89
16.0	1.09	0.09	1.00	0.65	0.89	0.77
32.0	0.46	0.07	0.41	0.37	0.45	0.41
64.0	0.14	0.01	0.13	0.15	0.16	0.16
128.0	k0.07	0.01	0.05	0.04	0.04	0.04

VES 8

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	10.58	0.42	10.16	7.22	6.95	7.09
1.0	3.62	0.43	3.21	3.00	2.54	2.77
2.0	1.56	0.13	1.47	1.26	1.22	1.24
4.0	1.35	0.07	1.25	0.74	0.97	0.86
8.0	1.14	0.08	1.06	0.66	0.82	0.74
16.0	0.91	0.08	0.85	0.53	0.64	0.59
32.0	1.47	0.04	1.39	0.37	0.35	0.36
64.0	0.23	0.00	0.23	0.19	0.18	0.19
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 9

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	3.41	0.15	3.29	2.62	2.09	2.36
1.0	1.51	0.07	1.41	1.06	1.23	1.15
2.0	0.91	0.05	0.89	0.58	0.68	0.63
4.0	0.90	0.04	0.84	0.54	0.55	0.55
8.0	0.86	0.11	0.78	0.54	0.57	0.56
16.0	0.69	0.06	0.60	0.48	0.46	0.47
32.0	0.47	0.03	0.44	0.30	0.37	0.34
64.0	0.11	0.01	0.10	0.12	0.10	0.11
128.0	0.10	0.00	0.10	0.06	0.04	0.05

VES 10

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.37	0.14	6.15	1.42	6.53	3.98
1.0	3.54	0.14	3.31	2.58	1.63	2.11
2.0	2.35	0.17	2.26	1.35	1.74	1.55
4.0	2.06	0.09	1.95	1.17	1.46	1.32
8.0	1.80	0.11	1.68	0.96	1.31	1.14
16.0	1.27	0.11	1.18	0.70	1.07	0.89
32.0	0.60	0.09	0.55	0.36	0.53	0.45
64.0	0.19	0.03	0.17	0.21	0.2	0.21
128.0	0.03	0.00	0.03	0.02	0.03	0.03

VES 11

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	14.44	0.07	14.28	7.92	9.24	11.24
1.0	15.03	0.15	14.99	2.27	16.26	9.27
2.60	0.21	2.39	2.30	2.04	2.17	
4.0	1.60	0.10	1.43	0.94	1.24	1.09
8.0	1.34	0.02	1.32	0.87	0.83	0.85
16.0	0.85	0.08	0.77	0.56	0.57	0.57
32.0	k0.59	0.00	0.59	0.40	0.40	0.40
64.0	0.17	0.02	0.15	0.12	0.09	0.11
128.0	0.11	0.05	0.06	0.03	0.09	0.06

VES 12

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.36	0.34	6.06	3.26	6.21	4.74
2.92	0.18	2.66	1.75	2.58	2.17	
2.0	1.66	0.10	1.57	1.20	1.12	1.16
4.0	1.37	0.08	1.24	0.96	0.84	0.90
8.0	1.19	0.01	1.18	0.88	0.69	0.79
16.0	0.88	0.07	0.77	0.58	0.51	0.55
32.0	0.36	0.02	0.33	0.31	0.30	0.31
64.0	0.05	0.00	0.05	0.06	0.07	0.07
128.0	0.13	0.05	0.08	0.07	0.07	0.07

Appendix E2. Continued

VES 13

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	4.32	0.19	3.94	3.89	2.41	3.15
1.0	1.53	0.12	1.41	1.46	1.02	1.24
2.0	1.18	0.07	1.12	0.80	0.74	0.77
4.0	0.90	0.05	0.84	0.62	0.56	0.59
8.0	0.88	0.05	0.81	0.57	0.50	0.54
16.0	0.62	0.05	0.55	0.46	0.39	0.43
32.0	0.41	0.05	0.37	0.32	0.29	0.31
64.0	0.15	0.01	0.14	0.11	0.11	0.11
128.0	--	--	--	--	--	--

VES 14

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	12.57	0.66	12.01	10.43	5.32	7.88
1.0	14.49	0.58	14.16	7.59	9.08	8.34
2.0	7.88	0.75	7.06	6.03	4.79	5.41
4.0	6.73	0.68	6.00	4.63	4.39	4.51
8.0	4.39	0.45	3.83	3.10	3.08	3.09
16.0	2.00	0.21	1.87	1.54	1.40	1.47
32.0	0.56	0.01	0.53	0.53	0.58	0.56
64.0	3.19	0.02	3.01	0.25	0.29	0.27
128.0	0.77	0.01	0.76	0.20	0.31	0.26

VES 15

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	6.23	0.67	5.36	1.72	11.95	6.84
1.0	3.62	0.17	3.48	2.50	2.67	2.59
2.0	1.51	0.05	1.50	0.89	1.30	1.09
4.0	1.17	0.13	1.03	0.71	0.81	0.76
8.0	0.96	0.06	0.89	0.64	0.61	0.63
16.0	0.70	0.04	0.64	0.49	0.51	0.50
32.0	0.41	0.02	0.38	0.31	0.29	0.30
64.0	0.14	0.00	0.14	0.14	0.13	0.14
128.0	0.07	0.00	0.07	0.04	0.02	0.03

VES 16

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	18.00	0.06	17.31	16.47	2.11	9.29
1.0	--	--	--	--	--	--
2.0	2.97	2.31	2.87	2.12	1.18	1.65
4.0	--	--	--	--	--	--
8.0	--	--	--	--	--	--
16.0	--	--	--	--	--	--
32.0	--	--	--	--	--	--
64.0	--	--	--	--	--	--
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 17

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	15.98	1.16	15.01	16.12	5.78	10.95
1.0	10.06	0.97	9.17	7.19	7.94	7.57
2.0	10.50	0.52	9.82	6.43	6.94	6.69
4.0	8.07	0.43	7.77	5.35	5.24	5.29
8.0	5.02	0.60	4.40	3.17	3.33	3.25
16.0	2.78	0.16	2.91	2.05	2.04	2.05
32.0	1.22	0.14	1.20	0.87	0.88	0.88
64.0	0.41	0.17	0.34	0.27	0.27	0.27
128.0	0.17	0.04	0.20	0.15	0.02	0.09

VES 18

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	3.49	0.42	3.26	2.35	4.07	3.21
1.0	4.41	0.18	4.13	2.78	2.51	2.64
2.0	2.57	0.21	2.35	2.14	1.26	1.70
4.0	2.74	0.29	2.64	1.98	1.51	1.74
8.0	2.11	0.40	2.11	1.68	1.48	1.58
16.0	1.13	0.63	1.02	0.91	1.03	0.97
32.0	0.90	0.03	0.87	0.67	0.66	0.66
64.0	0.96	0.97	0.91	0.85	1.06	0.96
128.0	0.09	0.06	0.14	0.09	0.02	0.06

VES 19

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.47	0.44	9.02	6.38	6.99	6.69
1.0	5.86	0.34	5.52	3.73	4.29	4.01
2.0	4.19	0.29	3.90	2.60	3.21	2.91
4.0	3.16	0.24	2.90	1.87	2.54	2.21
8.0	2.32	0.10	2.26	1.14	1.88	1.51
16.0	1.43	0.02	1.42	0.59	1.31	0.95
32.0	0.43	0.04	0.41	0.27	0.58	0.43
64.0	K0.09	0.01	0.09	0.07	0.21	0.14
128.0	0.08	0.00	0.08	0.05	0.07	0.06

VES 20

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	12.55	0.20	12.38	12.57	2.84	7.71
1.0	5.53	0.46	5.09	3.56	3.84	3.70
2.0	6.70	0.39	6.37	4.33	4.05	4.19
4.0	5.12	0.42	4.73	3.75	3.43	3.59
8.0	3.29	0.40	2.95	2.08	3.02	2.55
16.0	1.67	0.01	1.66	1.19	1.43	1.31
32.0	0.67	0.05	0.63	0.42	0.59	0.51
64.0	0.13	0.00	0.13	0.11	0.11	0.11
128.0	0.05	0.88	0.04	0.02	0.01	0.01

Appendix E2. Continued

VES 21

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	13.86	0.85	13.42	11.75	12.23	11.99
1.0	5.02	0.15	4.68	3.76	3.58	3.67
2.0	1.97	0.17	1.89	1.37	0.97	1.17
4.0	1.33	0.06	1.26	0.91	0.66	0.79
8.0	1.02	0.09	0.98	0.67	0.55	0.61
16.0	0.84	0.03	0.83	0.49	0.46	0.48
32.0	0.49	0.01	0.48	0.35	0.32	0.34
64.0	0.13	0.01	0.12	0.10	0.11	0.11
128.0	---	---	---	---	---	---

VES 22

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	17.27	0.36	17.16	8.32	6.71	7.52
1.0	4.29	0.19	4.07	2.89	3.07	2.98
2.0	3.29	0.22	3.11	2.02	2.19	2.11
4.0	3.02	0.24	2.79	1.86	1.90	1.88
8.0	1.94	0.13	1.81	1.29	1.65	1.47
16.0	1.33	0.04	1.23	0.99	0.99	0.99
32.0	0.58	0.04	0.54	0.50	0.51	0.51
64.0	0.16	0.00	0.15	0.11	0.12	0.12
128.0	0.06	0.00	0.06	0.03	0.03	0.03

VES 23

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	115.30	1.13	114.00	94.70	105.10	99.90
1.0	28.30	0.36	27.80	14.28	28.50	21.39
2.0	3.18	0.21	3.00	2.46	2.28	2.37
4.0	1.84	0.09	1.71	1.17	1.39	1.28
8.0	1.64	0.16	1.56	1.04	1.15	1.09
16.0	0.85	0.02	0.83	0.56	0.68	0.62
32.0	0.48	0.03	0.44	0.40	0.37	0.39
64.0	0.18	0.00	0.17	0.09	0.12	0.11
128.0	0.04	0.00	0.04	0.03	0.04	0.04

VES 24

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.87	0.04	9.77	7.92	3.83	5.88
1.0	4.44	0.36	4.25	2.13	4.07	3.10
2.0	4.01	0.29	3.72	2.61	2.58	2.60
4.0	3.50	0.26	3.23	2.58	2.03	2.31
8.0	2.83	0.12	2.62	2.30	1.52	1.91
16.0	1.27	0.21	1.94	1.06	0.97	1.02
32.0	0.61	0.09	0.52	0.57	0.18	0.38
64.0	0.18	0.01	0.16	0.10	0.08	0.09
128.0	0.09	0.00	0.09	0.05	0.02	0.04

Appendix E2. Continued

VES 25

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	7.84	0.71	7.19	6.01	6.59	6.30
1.0	5.02	0.23	4.78	2.72	2.13	2.43
2.0	1.88	0.04	1.75	1.43	1.36	1.40
4.0	1.38	0.06	1.33	0.95	0.85	0.90
8.0	0.91	0.13	0.75	0.59	0.67	0.63
16.0	0.84	0.08	0.78	0.53	0.50	0.52
32.0	0.35	0.02	0.33	0.27	0.22	0.25
64.0	0.19	0.03	0.16	0.06	0.05	0.06
128.0	--	--	--	--	--	--

VES 26

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	6.28	0.86	5.45	5.57	4.77	5.17
1.0	6.27	0.32	6.18	4.47	4.52	4.50
2.0	4.95	0.33	4.50	3.54	3.21	3.38
4.0	3.45	0.24	3.19	2.81	2.07	2.44
8.0	2.10	0.17	1.97	1.78	1.39	1.59
16.0	1.26	0.05	1.17	0.85	0.99	0.92
32.0	0.45	0.02	0.43	0.39	0.32	0.36
64.0	0.15	0.01	0.14	0.12	0.12	0.12
128.0	--	--	--	--	--	--

VES 28

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	4.82	0.08	4.78	3.75	2.51	3.13
1.0	2.75	0.15	2.58	2.22	1.64	1.93
2.0	2.02	0.15	1.96	1.51	1.26	1.39
4.0	1.67	0.09	1.50	1.22	0.92	1.07
8.0	1.33	0.07	1.28	0.99	0.77	0.88
16.0	0.75	0.09	0.63	0.71	0.48	0.59
32.0	0.40	0.05	0.35	0.33	0.22	0.28
64.0	0.19	0.05	0.14	0.11	0.13	0.12
128.0	0.03	0.01	0.02	0.02	0.03	0.03

VES 29

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	26.90	1.20	25.80	16.31	19.13	17.72
1.0	13.32	0.93	12.41	10.33	9.79	10.06
2.0	7.31	0.36	6.99	5.56	5.07	5.32
4.0	3.32	0.19	3.13	2.45	2.41	2.43
8.0	1.77	0.14	1.62	1.35	1.25	1.30
16.0	0.87	0.06	0.80	0.71	0.70	0.71
32.0	0.29	0.03	0.28	0.26	0.26	0.26
64.0	0.09	0.00	0.09	0.07	0.07	0.07
128.0	0.06	0.00	0.06	0.01	0.01	0.01

Appendix E2. Continued

VES 29

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	4.76	0.32	4.53	3.03	3.79	3.41
1.0	2.78	0.15	2.82	2.03	1.48	1.76
2.0	1.96	0.14	1.88	1.28	1.34	1.31
4.0	2.17	0.13	2.07	1.29	1.33	1.31
8.0	1.87	0.19	1.66	1.18	1.23	1.21
16.0	1.28	0.15	1.12	0.90	0.91	0.91
32.0	0.59	0.03	0.57	0.51	0.50	0.51
64.0	0.11	0.01	0.10	0.16	0.15	0.16
128.0	0.03	0.00	0.03	0.02	0.02	0.02

VES 30

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	11.09	0.49	10.60	7.66	7.20	7.43
1.0	9.46	0.53	8.94	5.42	6.62	6.02
2.0	10.06	0.51	9.57	5.64	5.74	5.69
4.0	8.55	0.71	7.80	4.80	4.73	4.77
8.0	4.43	0.57	3.86	3.54	3.44	3.49
16.0	1.92	0.24	1.70	1.36	2.17	1.77
32.0	0.74	0.13	0.61	0.50	0.51	0.51
64.0	0.42	0.01	0.41	0.05	0.05	0.05
128.0	0.36	0.06	0.28	0.09	0.05	0.07

VES 31

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.42	0.29	5.18	4.41	3.52	3.97
1.0	3.16	0.12	3.01	2.40	1.91	2.16
2.0	1.95	0.14	1.90	1.41	1.32	1.37
4.0	1.49	0.09	1.32	0.81	0.93	0.87
8.0	1.24	0.10	1.17	0.84	0.83	0.84
16.0	0.84	0.04	0.72	0.62	0.57	0.60
32.0	0.37	0.02	0.39	0.37	0.39	0.38
64.0	0.15	0.01	0.13	0.10	0.20	0.15
128.0	--	--	--	--	--	--

VES 32

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	11.48	0.57	10.96	4.33	13.96	9.15
1.0	5.06	0.32	4.72	3.28	4.62	3.95
2.0	2.63	0.24	2.48	1.66	2.13	1.90
4.0	1.85	0.15	1.77	1.17	1.39	1.28
8.0	1.49	0.15	1.37	0.87	1.11	0.99
16.0	0.96	0.05	0.91	0.64	0.73	0.69
32.0	0.38	0.05	0.34	0.31	0.36	0.34
64.0	0.11	0.01	0.10	0.11	0.11	0.11
128.0	0.04	0.00	0.04	0.02	0.02	0.02

Appendix E2. Continued

VES 33

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	3.67	0.22	3.47	3.87	1.81	2.84
1.0	2.06	0.11	1.96	1.67	1.17	1.42
2.0	1.67	0.09	1.56	1.13	0.93	1.03
4.0	1.55	0.03	1.44	1.08	0.86	0.97
8.0	1.49	0.16	1.40	1.03	0.86	0.94
16.0	1.09	0.09	0.94	0.65	0.66	0.66
32.0	0.38	0.04	0.35	0.31	0.30	0.31
64.0	0.09	0.01	0.08	0.08	0.08	0.08
128.0	0.04	0.00	0.03	0.03	0.03	0.03

VES 34

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	81.50	5.73	75.40	67.90	64.20	66.05
1.0	36.90	1.60	35.20	24.20	31.60	27.90
2.0	8.98	0.51	8.49	8.95	6.36	7.66
4.0	4.42	0.15	4.18	2.85	3.28	3.07
8.0	2.80	0.28	2.52	1.93	1.89	1.91
16.0	1.32	0.09	1.21	1.22	0.88	1.05
32.0	0.43	0.00	0.44	0.33	0.47	0.40
64.0	0.26	0.04	0.22	0.28	0.21	0.25
128.0	0.16	0.04	0.12	0.09	0.09	0.09

VES 35

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	39.70	1.46	38.20	25.40	24.20	24.80
1.0	26.90	2.11	24.80	17.50	19.48	18.49
2.0	25.00	1.15	23.90	15.16	17.65	16.41
4.0	14.56	0.99	13.57	10.76	11.44	11.10
8.0	6.41	0.46	5.98	5.12	5.53	5.33
16.0	2.22	0.09	2.12	1.88	1.85	1.87
32.0	0.47	0.03	0.43	0.38	0.47	0.43
64.0	0.13	0.10	0.03	0.07	0.11	0.09
128.0	0.13	0.02	0.11	0.05	0.00	0.03

VES 36

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	14.36	1.02	13.35	6.50	13.82	10.16
1.0	15.90	0.60	15.30	8.45	10.86	9.66
2.0	14.81	0.87	13.90	7.67	9.13	8.40
4.0	11.52	1.42	10.12	7.26	7.27	7.27
8.0	7.08	0.67	6.46	5.26	5.24	5.25
16.0	3.36	0.15	3.20	2.27	2.62	2.45
32.0	0.57	0.01	0.56	0.46	0.60	0.53
64.0	0.08	0.01	0.07	0.06	0.07	0.07
128.0	0.04	0.00	0.04	0.02	0.01	0.02

Appendix E2. Continued

VES 37

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	18.40	1.32	17.11	6.97	19.63	13.30
1.0	20.20	2.41	17.80	12.87	12.69	12.78
2.0	22.40	0.98	21.50	11.69	11.68	11.69
4.0	15.96	1.26	14.67	8.36	8.21	8.29
8.0	10.96	0.71	10.25	5.55	5.59	5.57
16.0	2.12	0.21	1.91	2.64	2.07	2.36
32.0	0.63	0.03	0.61	0.59	0.47	0.53
64.0	0.11	0.05	0.06	0.04	0.05	0.05
128.0	0.07	0.01	0.06	0.06	0.06	0.06

VES 38

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	1420.00	108.70	1309.00	1119.00	1003.00	1061.00
1.0	858.00	64.10	793.00	637.00	647.00	642.00
2.0	425.00	28.30	396.11	352.00	317.00	334.50
4.0	131.50	4.76	126.00	106.90	119.00	111.95
8.0	21.20	0.64	21.40	20.90	15.29	18.10
16.0	3.81	0.15	3.56	2.19	3.39	2.79
32.0	0.74	0.04	0.68	0.64	0.36	0.50
64.0	0.10	0.04	0.06	0.04	0.10	0.07
128.0	0.09	0.03	0.06	0.02	0.03	0.03

VES 39

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	8.34	0.58	7.67	6.32	6.33	6.33
1.0	6.30	0.19	6.01	4.23	3.88	4.06
2.0	2.47	0.18	2.31	1.78	2.05	1.92
4.0	1.57	0.11	1.49	1.01	1.17	1.09
8.0	1.18	0.11	1.11	0.85	0.83	0.84
16.0	0.86	0.04	0.79	0.56	0.59	0.58
32.0	0.45	0.04	0.40	0.35	0.32	0.34
64.0	0.12	0.01	0.10	0.07	0.13	0.10
128.0	0.04	0.00	0.04	0.03	0.01	0.02

VES 40

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.99	0.16	5.85	4.34	3.21	3.78
1.0	2.39	0.15	2.23	1.49	1.90	1.70
2.0	2.11	0.13	2.00	1.33	1.38	1.36
4.0	1.92	0.12	1.80	1.22	1.27	1.25
8.0	1.69	0.12	1.54	1.07	1.11	1.09
16.0	1.02	0.09	0.87	0.69	0.77	0.73
32.0	0.42	0.04	0.41	0.40	0.31	0.36
64.0	0.02	0.01	0.03	0.01	0.05	0.03
128.0	0.07	0.04	0.06	0.05	0.08	0.07

Appendix E2. Continued

VES 41

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	6.49	0.67	5.83	6.49	4.82	5.66
1.0	5.36	0.19	5.16	3.90	3.26	3.58
2.0	2.73	0.16	2.62	2.06	1.80	1.93
4.0	1.93	0.13	1.80	1.46	1.24	1.35
8.0	1.36	0.09	1.26	1.00	0.92	0.96
16.0	0.91	0.07	0.85	0.80	0.53	0.67
32.0	0.42	0.00	0.41	0.32	0.30	0.31
64.0	0.14	0.07	0.07	0.07	0.06	0.07
128.0	0.06	0.01	0.05	0.04	0.02	0.03

VES42

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	18.00	1.01	16.49	8.76	14.99	11.88
1.0	13.91	0.91	13.04	6.67	11.98	9.33
2.0	16.90	0.47	16.36	8.67	9.09	8.88
4.0	11.65	0.99	10.52	7.19	7.24	7.22
8.0	8.37	0.84	7.52	4.67	4.71	4.69
16.0	3.30	0.25	3.09	2.24	2.43	2.34
32.0	0.12	0.07	0.06	0.60	0.60	0.60
64.0	0.11	0.09	0.02	0.09	0.09	0.09
128.0	--	--	--	--	--	--

VES 43

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	56.90	2.39	54.60	41.30	39.70	40.50
1.0	35.80	1.75	34.10	21.90	27.20	24.55
2.0	20.70	0.44	19.20	13.52	15.29	14.41
4.0	11.33	0.34	10.94	7.94	6.57	7.26
8.0	3.94	0.16	3.81	2.95	2.59	2.77
16.0	1.87	0.07	1.83	0.99	1.66	1.33
32.0	0.35	0.05	0.30	0.35	0.22	0.29
64.0	0.12	0.01	0.10	0.12	0.06	0.09
128.0	0.05	0.00	0.04	0.01	0.01	0.01

VES 44

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	51.40	2.30	48.60	28.40	37.80	33.10
1.0	36.40	1.60	33.80	23.90	25.30	24.60
2.0	20.20	0.90	18.40	12.03	16.85	14.44
4.0	8.66	0.54	7.99	5.31	7.67	6.49
8.0	4.50	0.33	4.17	2.57	4.21	3.39
16.0	2.35	0.06	2.29	1.08	2.17	1.63
32.0	0.45	0.02	0.44	0.32	0.47	0.40
64.0	0.08	0.01	0.07	0.06	0.06	0.06
128.0	0.04	0.00	0.04	0.01	0.00	0.01

Appendix E2. Continued

VES 45

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	11.18	0.42	10.77	8.61	7.01	7.81
1.0	3.48	0.20	3.28	2.90	2.69	2.80
2.0	1.84	0.10	1.72	1.33	1.22	1.28
4.0	1.55	0.07	1.49	0.97	1.01	0.99
8.0	1.29	0.10	1.18	0.80	0.94	0.87
16.0	1.05	0.06	0.99	0.65	0.77	0.71
32.0	0.50	0.02	0.48	0.31	0.47	0.39
64.0	0.14	0.00	0.12	0.09	0.12	0.11
128.0	0.14	0.03	0.11	0.03	0.03	0.03

VES 46

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.14	0.27	5.87	4.35	4.43	4.39
1.0	3.73	0.18	3.49	2.37	2.61	2.49
2.0	2.44	0.51	1.95	1.30	1.92	1.61
4.0	2.58	0.20	2.38	1.76	1.80	1.78
8.0	1.93	0.11	1.82	1.19	1.33	1.26
16.0	0.98	0.09	0.85	0.73	0.86	0.80
32.0	0.38	0.02	0.37	0.26	0.39	0.33
64.0	0.10	0.02	0.07	0.05	0.05	0.05
128.0	0.04	0.02	0.03	0.04	0.04	0.04

VES 47

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	101.50	6.70	94.90	80.60	90.30	85.45
1.0	60.30	2.80	57.70	27.90	57.30	42.60
2.0	19.58	0.91	18.69	9.37	21.00	15.19
4.0	3.55	0.23	3.33	3.13	3.14	3.14
8.0	2.04	0.06	1.99	0.99	1.78	1.39
16.0	0.93	0.04	0.89	0.54	0.83	0.69
32.0	0.32	0.01	0.30	0.23	0.27	0.25
64.0	0.09	0.01	0.09	0.07	0.05	0.06
128.0	0.04	0.00	0.03	0.02	0.01	0.02

VES 48

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	61.30	2.50	58.80	40.70	39.00	39.85
1.0	46.20	2.64	43.60	27.20	26.50	26.85
2.0	38.80	1.71	37.20	21.70	16.19	18.95
4.0	15.51	0.93	14.58	14.66	9.04	11.85
8.0	8.66	0.15	8.60	7.62	3.53	5.58
16.0	2.88	0.16	2.69	1.77	1.82	1.80
32.0	0.44	0.05	0.40	0.44	0.40	0.42
64.0	0.27	0.03	0.24	0.12	0.06	0.09
128.0	0.32	0.07	0.25	0.12	0.43	0.28

Appendix E2. Continued

VES 49

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	34.90	1.46	33.40	26.30	20.40	23.35
1.0	21.10	1.14	19.96	14.03	15.66	14.85
2.0	15.62	0.59	15.00	10.14	11.10	10.62
4.0	7.60	0.45	6.99	5.42	5.66	5.54
8.0	5.41	0.45	5.08	2.46	4.97	3.72
16.0	1.77	0.10	1.73	1.62	1.20	1.41
32.0	0.51	0.03	0.34	0.17	0.34	0.26
64.0	0.08	0.01	0.16	0.15	0.08	0.12
128.0	--	--	--	--	--	--

VES 50

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	18.13	2.03	17.13	9.43	17.61	13.52
1.0	12.90	0.66	12.27	6.43	11.25	8.84
2.0	7.36	0.40	7.00	4.71	5.69	5.20
4.0	4.06	0.31	3.73	2.74	3.22	2.98
8.0	2.69	0.15	2.57	1.96	1.80	1.88
16.0	1.41	0.06	1.29	1.08	0.96	1.02
32.0	0.24	0.01	0.24	0.28	0.28	0.28
64.0	0.05	0.01	0.04	0.02	0.05	0.04
128.0	--	--	--	--	--	--

VES 51

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	17.73	1.82	15.84	19.92	13.99	16.96
1.0	9.60	0.26	9.31	8.02	5.18	6.60
2.0	4.76	0.29	4.48	3.44	2.92	3.18
4.0	4.37	0.27	4.10	3.32	2.34	2.83
8.0	2.70	0.29	2.47	1.96	2.02	1.99
16.0	1.37	0.09	1.24	1.28	0.92	1.10
32.0	0.51	0.04	0.46	0.46	0.39	0.43
64.0	0.10	0.01	0.09	0.09	0.11	0.10
128.0	0.03	0.00	0.02	0.02	0.02	0.02

VES 52

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.37	0.52	8.86	8.65	5.34	6.99
1.0	5.16	0.25	4.91	3.94	3.53	3.74
2.0	2.35	0.14	2.22	1.81	1.76	1.79
4.0	1.06	0.05	1.00	0.71	0.75	0.73
8.0	0.79	0.06	0.69	0.52	0.39	0.46
16.0	0.65	0.04	0.59	0.42	0.42	0.42
32.0	0.37	0.01	0.36	0930	0.26	0.28
64.0	0.16	0.01	0.14	0.12	0.11	0.12
128.0	0.07	0.01	0.06	0.03	0.02	0.03

Appendix E2. Continued

VES 53

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	21.00	1.91	19.31	15.35	16.11	15.73
1.0	12.13	0.77	11.34	10.01	8.25	9.13
2.0	3.24	0.18	3.08	2.74	2.94	2.84
4.0	1.00	0.04	0.93	0.69	0.72	0.71
8.0	0.78	0.06	0.70	0.46	0.42	0.44
16.0	0.56	0.03	0.55	0.43	0.35	0.39
32.0	0.29	0.01	0.29	0.24	0.13	0.19
64.0	0.06	0.00	0.06	0.06	0.06	0.06
128.0	0.04	0.01	0.03	0.02	0.01	0.02

VES 54

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	17.67	1.32	16.33	11.29	14.40	12.85
1.0	14.39	0.983	13.48	9.86	9.72	9.79
2.0	10.73	0.69	10.12	7.98	6.76	7.37
4.0	8.06	0.31	7.79	6.23	4.51	5.37
8.0	5.79	0.28	5.47	4.70	2.92	3.81
16.0	1.83	0.16	1.81	1.89	1.13	1.51
32.0	0.55	0.01	0.43	0.40	0.36	0.38
64.0	0.04	0.01	0.04	0.02	0.02	0.02
128.0	0.03	0.00	0.03	0.02	0.03	0.03

VES 55

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	13.06	0.63	12.44	7.64	10.01	8.83
1.0	8.47	0.46	8.01	5.49	5.91	5.70
2.0	6.94	0.39	6.56	3.94	5.21	4.58
4.0	5.70	0.26	5.40	3.36	4.14	3.75
8.0	2.89	0.29	2.51	1.94	2.72	2.33
16.0	2.00	0.17	1.91	1.39	1.45	1.42
32.0	0.35	0.03	0.32	0.39	0.47	0.43
64.0	0.16	0.02	0.14	0.04	0.02	0.03
128.0	0.13	0.03	0.10	0.12	0.13	0.13

VES 56

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	30.90	1.32	29.60	24.70	15.36	20.03
1.0	27.50	1.11	26.40	23.40	11.31	17.36
2.0	21.30	0.63	20.80	19.43	6.84	13.14
4.0	11.93	0.70	11.26	11.74	3.64	7.69
8.0	7.45	0.35	7.10	6.61	3.63	5.12
16.0	2.37	0.20	2.07	2.57	1.95	2.26
32.0	0.88	0.00	0.88	0.72	0.63	0.68
64.0	0.06	0.02	0.06	0.06	0.06	0.06
128.0	0.04	0.02	0.04	0.02	0.01	0.02

Appendix E2. Continued

VES 57

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	42.60	1.66	42.10	22.10	35.10	28.60
1.0	30.80	1.37	29.40	21.00	19.29	20.15
2.0	23.90	1.54	22.40	15.29	16.38	15.84
4.0	10.82	1.62	9.17	10.05	8.07	9.06
8.0	9.70	0.32	9.45	5.63	6.58	6.11
16.0	2.23	0.37	1.83	2.26	2.27	2.27
32.0	0.78	0.01	0.77	0.71	0.71	0.71
64.0	0.09	0.01	0.08	0.03	0.06	0.05
128.0	--	--	--	--	--	--

VES 58

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	61.20	3.89	57.30	37.20	47.00	42.10
1.0	49.70	2.66	47.10	33.00	34.20	33.60
2.0	31.10	1.64	29.40	22.90	21.50	22.20
4.0	19.38	1.06	18.33	13.19	13.71	13.45
8.0	10.44	0.62	9.81	7.07	8.27	7.67
16.0	2.83	0.18	2.69	2.40	2.68	2.54
32.0	0.37	0.01	0.36	0.35	0.42	0.39
64.0	0.02	0.01	0.02	0.00	0.13	0.07
128.0	--	--	--	--	--	--

VES 59

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	42.90	1.82	41.00	26.70	29.50	28.10
1.0	33.10	1.04	32.10	17.51	24.20	20.86
2.0	27.20	1.18	26.00	13.73	20.70	17.22
4.0	18.57	1.47	17.21	11.42	13.96	12.69
8.0	8.73	0.99	7.77	5.31	9.56	7.44
16.0	3.24	0.26	3.10	2.62	3.01	2.82
32.0	0.84	0.05	0.79	0.64	0.72	0.68
64.0	0.22	0.01	0.21	0.11	0.05	0.08
128.0	K0.06	0.01	0.05	0.02	0.02	0.02

VES 60

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	36.60	1.74	34.80	25.90	31.30	28.60
1.0	17.11	0.53	16.58	12.08	12.53	12.31
2.0	5.98	0.31	5.67	4.01	4.42	4.22
4.0	3.46	0.24	3.22	2.25	2.44	2.35
8.0	2.33	0.21	2.13	1.61	1.76	1.68
16.0	1.33	0.09	1.24	0.06	1.00	1.03
32.0	0.39	0.01	0.38	0.35	0.33	0.34
64.0	0.16	0.01	0.16	0.07	0.01	0.04
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 61

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.36	0.57	8.83	8.58	6.39	7.49
1.0	4.01	0.26	3.75	3.23	2.98	3.11
2.0	2.57	0.15	2.54	1.79	1.76	1.78
4.0	1.56	0.11	1.45	1.01	1.19	1.10
8.0	0.92	0.07	0.86	0.57	0.73	0.65
16.0	0.39	0.03	0.36	0.27	0.35	0.31
32.0	0.17	0.02	0.16	0.12	0.14	0.13
64.0	0.06	0.01	0.05	0.04	0.04	0.04
128.0	0.03	0.01	0.02	0.02	0.02	0.02

VES 62

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	10.12	0.67	9.45	6.92	8.19	7.56
1.0	7.00	0.46	6.58	4.94	5.01	4.98
2.0	5.20	0.36	4.85	3.60	3.64	3.62
4.0	3.12	0.25	2.84	2.39	2.31	2.35
8.0	1.57	0.11	1.41	1.32	1.19	1.26
16.0	0.55	0.04	0.52	0.45	0.49	0.47
32.0	0.23	0.02	0.21	0.20	0.13	0.17
64.0	0.06	0.01	0.05	0.05	0.04	0.05
128.0	0.03	0.01	0.03	0.02	0.02	0.02

VES 63

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	7.87	0.57	7.48	6.85	5.51	6.18
1.0	4.61	0.31	4.29	3.44	3.11	3.28
2.0	3.64	0.21	3.60	2.71	2.05	2.38
4.0	2.73	0.19	2.60	2.07	1.59	1.83
8.0	2.45	0.15	2.27	1.73	1.47	1.60
16.0	1.14	0.14	1.01	1.13	0.76	0.95
32.0	0.55	0.03	0.52	0.39	0.48	0.44
64.0	0.27	0.14	0.13	0.12	0.16	0.14
128.0	--	--	--	--	--	--

VES 64

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.71	0.28	5.46	4.06	4.22	4.14
1.0	4.18	0.19	4.00	2.94	2.51	2.73
2.0	3.92	0.25	3.68	2.66	2.40	2.53
4.0	3.63	0.27	3.35	2.48	2.37	2.43
8.0	2.84	0.24	2.69	2.07	2.04	2.06
16.0	1.49	0.14	1.32	1.16	1.30	1.23
32.0	0.57	0.02	0.55	0.49	0.45	0.47
64.0	0.09	0.01	0.09	0.09	0.08	0.09
128.0	0.04	0.01	0.04	0.02	0.02	0.02

Appendix E2. Continued

VES 65

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	29.80	2.12	27.70	21.80	22.10	21.95
1.0	17.88	1.15	16.73	13.33	12.60	12.97
2.0	10.08	0.66	9.44	7.60	7.07	7.34
4.0	5.65	0.39	5.24	4.30	4.26	4.28
8.0	3.31	0.21	3.00	2.31	2.48	2.40
16.0	1.46	0.13	1.31	1.25	1.14	1.19
32.0	0.57	0.03	0.56	0.44	0.49	0.47
64.0	0.11	0.01	0.10	0.07	0.06	0.07
128.0	0.05	0.01	0.04	0.03	0.03	0.03

VES 66

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	29.70	1.88	27.80	24.10	21.10	22.60
1.0	15.90	0.91	15.01	11.12	12.29	11.71
2.0	9.39	0.55	9.02	6.51	7.18	6.85
4.0	5.46	0.36	5.11	3.49	4.32	3.91
8.0	3.45	0.24	3.16	2.19	2.61	2.40
16.0	1.69	0.11	1.59	1.26	1.37	1.32
32.0	0.53	0.03	0.50	0.30	0.58	0.44
64.0	0.07	0.00	0.07	0.04	0.07	0.06
128.0	0.0-4	0.02	0.02	0.02	0.03	0.03

VES 67

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	4.57	0.25	4.33	3.13	3.32	3.23
1.0	3.42	0.16	3.23	2.18	2.29	2.24
2.0	3.02	0.16	2.86	1.84	2.03	1.94
4.0	2.21	0.19	2.04	1.59	1.43	1.51
8.0	1.85	0.18	1.66	1.19	1.38	1.29
16.0	1.35	0.07	1.27	0.95	0.93	0.94
32.0	0.43	0.02	0.41	0.38	0.34	0.36
64.0	0.07	0.01	0.07	0.06	0.08	0.07
128.0	0.04	0.00	0.04	0.01	0.01	0.01

VES 68

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	4.98	0.28	4.77	3.96	3.79	3.88
1.0	2.36	0.11	2.20	1.85	1.67	1.76
2.0	1.36	0.09	1.19	0.84	0.96	0.90
4.0	0.91	0.08	0.83	0.63	0.85	0.74
8.0	0.73	0.95	0.67	0.43	0.54	0.49
16.0	0.54	0.08	0.46	0.35	0.49	0.42
32.0	0.33	0.02	0.31	0.27	0.28	0.28
64.0	0.14	0.03	0.11	0.14	0.10	0.12
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 69

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.11	0.64	14.53	10.10	9.51	19.81
1.0	11.37	0.63	10.74	7.92	6.86	7.39
2.0	10.66	0.63	10.03	7.64	5.93	6.79
4.0	8.54	0.63	7.88	6.53	5.03	5.78
8.0	4.87	0.41	4.49	4.59	2.99	3.79
16.0	1.59	0.09	1.51	1.74	1.20	1.47
32.0	0.46	0.01	0.43	0.44	0.28	0.36
64.0	0.07	0.00	0.07	0.05	0.05	0.05
128.0	0.03	0.01	0.02	0.02	0.02	0.02

VES 70

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.05	0.49	5.58	5.55	5.03	5.29
1.0	3.19	0.14	2.95	2.17	2.35	2.26
2.0	1.70	0.09	1.61	1.21	1.23	1.22
4.0	1.56	0.06	1.44	0.93	0.93	0.93
8.0	1.61	0.09	1.48	0.91	0.98	0.95
16.0	1.13	0.12	0.99	0.75	0.93	0.84
32.0	0.59	0.02	0.58	0.56	0.51	0.54
64.0	0.29	0.00	0.28	0.27	0.28	0.28
128.0	0.02	0.00	0.01	0.03	0.04	0.04

VES 71

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	21.00	1.17	19.95	13.02	15.05	14.04
1.0	18.35	0.96	17.33	10.91	12.77	11.84
2.0	12.80	1.03	11.90	8.79	9.50	9.15
4.0	8.87	0.68	8.22	6.23	6.81	6.52
8.0	4.26	0.36	3.93	3.73	3.40	3.57
16.0	1.89	0.07	1.78	1.40	1.62	1.51
32.0	0.46	0.02	0.43	0.35	0.33	0.34
64.0	0.05	0.00	0.05	0.05	0.05	0.05
128.0	0.04	0.02	0.03	0.02	0.02	0.02

VES 72

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	17.44	0.14	17.23	13.39	6.82	10.11
1.0	6.11	0.51	5.52	6.16	3.96	5.06
2.0	5.81	0.44	5.37	4.21	3.63	3.92
4.0	4.14	0.38	3.72	2.90	3.09	2.99
8.0	3.10	0.19	2.91	2.08	2.30	2.19
16.0	1.55	0.08	1.46	1.16	1.25	1.21
32.0	0.56	0.02	0.56	0.50	0.47	0.49
64.0	0.11	0.02	0.09	0.11	0.10	0.11
128.0	0.04	0.00	0.03	0.03	0.03	0.03

Appendix E2. Continued

VES 73

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	17.20	0.89	16.32	13.44	9.50	11.47
1.0	14.39	0.92	13.46	9.48	9.49	9.49
2.0	12.46	0.83	11.63	8.07	8.51	8.29
4.0	9.24	0.63	8.55	6.11	6.75	6.43
8.0	5.78	0.41	5.37	4.30	4.13	4.22
16.0	2.28	0.16	2.07	1.78	2.12	1.95
32.0	0.67	0.04	0.63	0.55	0.51	0.53
64.0	0.12	0.00	0.11	0.09	0.09	0.09
128.0	0.03	0.01	0.02	0.02	0.02	0.02

VES 74

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.69	0.84	4.80	4.09	3.98	4.04
1.0	3.89	0.29	3.89	2.85	2.83	2.84
2.0	3.64	0.33	3.47	2.32	2.34	2.33
4.0	3.97	0.38	3.58	2.28	2.27	2.28
8.0	3.33	0.19	3.14	2.06	2.05	2.06
16.0	2.34	0.15	2.21	1.43	1.34	1.39
32.0	0.88	0.11	0.76	0.59	0.60	0.60
64.0	0.19	0.01	0.18	0.15	0.15	0.15
128.0	--	--	--	--	--	--

VES 75

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	8.48	0.51	8.06	7.56	7.95	7.76
1.0	2.88	0.12	2.78	2.27	2.19	2.23
2.0	2.10	0.19	1.86	1.25	1.37	1.31
4.0	2.11	0.01	2.01	1.29	1.40	1.35
8.0	1.78	0.11	1.76	1.21	1.15	1.18
16.0	1.41	0.11	1.29	0.89	0.82	0.86
32.0	0.72	0.02	0.70	0.45	0.37	0.41
64.0	0.15	0.04	0.11	0.11	0.10	0.11
128.0	--	--	--	--	--	--

VES 76

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	12.25	0.79	11.48	10.34	8.10	9.22
1.0	8.75	0.36	8.39	6.59	5.11	5.85
2.0	6.41	0.36	6.06	5.28	3.20	4.24
4.0	4.90	0.21	4.68	3.93	2.44	3.19
8.0	2.15	0.25	1.93	2.28	2.26	2.27
16.0	1.56	0.05	1.49	1.32	1.33	1.33
32.0	0.43	0.02	0.41	0.44	0.43	0.44
64.0	0.05	0.00	0.05	0.07	0.07	0.07
128.0	0.05	0.01	0.04	0.02	0.02	0.02

Appendix E2. Continued

VES 77

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	21.60	1.24	20.50	14.47	14.17	14.32
1.0	16.33	1.09	15.16	11.26	11.00	11.13
2.0	9.96	0.86	9.11	8.24	6.83	7.54
4.0	3.84	0.29	3.54	3.37	3.57	3.47
8.0	2.55	0.12	2.48	2.14	1.39	1.77
16.0	1.12	0.06	1.09	0.91	0.84	0.88
32.0	0.54	0.04	0.48	0.46	0.39	0.43
64.0	0.48	0.01	0.47	0.46	0.49	0.48
128.0	0.15	0.00	0.15	0.17	0.14	0.16

VES 78

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	5.79	0.43	5.36	4.57	2.64	3.61
1.0	3.92	0.16	3.80	2.66	2.77	2.72
2.0	3.02	0.11	2.95	1.71	2.11	1.91
4.0	2.69	0.17	2.52	1.73	1.63	1.68
8.0	1.98	0.26	1.72	1.34	1.33	1.34
16.0	2.14	0.07	2.06	0.98	0.98	0.98
32.0	0.35	0.02	0.32	0.33	0.33	0.33
64.0	0.07	0.01	0.06	0.06	0.06	0.06
128.0	0.03	0.00	0.02	0.08	0.09	0.09

VES 79

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	5.34	0.37	5.02	4.28	3.68	3.98
1.0	2.17	0.15	2.00	1.74	1.67	1.71
2.0	1.07	0.02	0.98	0.81	0.75	0.78
4.0	0.66	0.04	0.61	0.39	0.42	0.41
8.0	0.77	0.07	0.69	0.38	0.34	0.36
16.0	0.68	0.00	0.67	0.53	0.53	0.53
32.0	0.51	0.04	0.47	0.37	0.38	0.38
64.0	0.27	0.01	0.26	0.23	0.25	0.24
128.0	0.12	0.02	0.10	0.06	0.06	0.06

VES 80

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	6.16	0.63	5.63	4.07	6.80	5.44
1.0	5.56	0.22	5.30	3.07	4.11	3.59
2.0	3.81	0.28	3.49	2.11	2.99	2.55
4.0	2.62	0.14	2.41	1.28	2.17	1.73
8.0	1.11	0.12	0.95	0.92	0.95	0.94
16.0	0.71	0.02	0.68	0.53	0.56	0.55
32.0	0.35	0.02	0.32	0.18	0.18	0.18
64.0	0.09	0.01	0.08	0.06	0.06	0.06
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 81

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	28.30	1.45	26.90	18.72	19.22	18.97
1.0	16.53	1.45	15.02	11.95	12.55	12.25
2.0	13.58	0.63	13.02	9.34	9.22	9.28
4.0	9.76	0.40	9.34	6.38	6.69	6.54
8.0	4.16	0.39	3.75	3.06	3.93	3.50
16.0	1.21	0.09	1.10	0.99	1.33	1.16
32.0	0.06	0.01	0.32	0.27	0.25	0.26
64.0	0.06	0.00	0.06	0.03	0.04	0.04
128.0	0.04	0.00	0.04	0.02	0.01	0.02

VES 82

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	2.76	0.36	2.49	2.09	2.47	2.28
1.0	3.30	0.17	3.16	2.11	2.12	2.12
2.0	2.45	0.18	2.27	1.58	1.67	1.63
4.0	1.97	0.11	1.85	1.11	1.42	1.27
8.0	1.21	0.13	1.09	0.87	0.92	0.90
16.0	0.99	0.04	0.96	0.66	0.65	0.66
32.0	0.32	0.03	0.30	0.21	0.23	0.22
64.0	0.06	0.00	0.06	0.06	0.07	0.07
128.0	0.05	0.01	0.05	0.02	0.01	0.02

VES 83

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.25	0.54	8.74	7.31	5.95	6.63
1.0	6.89	0.42	6.46	4.39	4.75	4.57
2.0	5.80	0.45	5.35	3.63	4.14	3.89
4.0	4.80	0.27	4.49	2.71	3.80	3.26
8.0	3.94	0.13	3.83	2.63	2.36	2.50
16.00	1.39	0.08	1.32	0.81	1.31	1.06
32.0	0.33	0.02	0.33	0.19	0.33	0.26
64.0	0.03	0.00	0.03	0.04	0.04	0.04
128.0	0.04	0.00	0.03	0.04	0.04	0.04

VES 84

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.82	0.31	6.20	5.58	2.78	4.18
1.0	3.42	0.22	3.20	2.41	2.08	2.25
2.0	2.30	0.07	2.16	1.61	1.59	1.60
4.0	1.72	0.10	1.61	1.16	1.12	1.14
8.0	1.61	0.08	1.52	0.95	0.90	0.93
16.0	0.54	0.01	0.53	0.47	0.64	0.56
32.0	0.29	0.01	0.28	0.22	0.23	0.23
64.0	0.03	0.00	0.03	0.05	0.05	0.05
128.0	0.10	0.02	0.08	0.06	0.06	0.06

Appendix E2. Continued

VES 85

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	22.50	1.82	20.70	17.02	14.76	15.89
1.0	18.12	1.24	16.87	13.31	11.78	12.55
2.0	13.55	0.75	12.83	10.26	8.18	9.22
4.0	8.94	0.46	8.51	6.60	6.12	6.36
8.0	3.46	0.44	2.95	3.51	3.57	3.54
16.0	1.91	0.07	1.84	1.79	1.78	1.79
32.0	0.53	0.00	0.50	0.46	0.46	0.46
64.0	0.08	0.00	0.08	0.05	0.05	0.05
128.0	--	--	--	--	--	--

VES 86

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	16.71	0.92	15.81	11.83	11.81	11.82
1.0	8.43	0.68	7.69	6.83	5.84	6.34
2.0	4.94	0.30	4.67	3.91	3.30	3.61
4.0	2.56	0.13	2.43	2.02	1.81	1.92
8.0	1.66	0.12	1.48	1.18	1.01	1.10
16.0	1.08	0.04	1.07	0.85	0.67	0.76
32.0	0.54	0.03	0.50	0.32	0.33	0.33
64.0	0.07	0.01	0.07	0.11	0.10	0.11
128.0	--	--	--	--	--	--

VES 87

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	6.78	0.60	6.28	5.60	5.16	5.39
1.0	3.18	0.26	2.92	2.50	2.70	2.60
2.0	2.54	0.17	2.37	1.49	1.49	1.49
4.0	1.62	0.10	1.47	1.17	1.27	1.22
8.0	1.32	0.16	1.14	0.89	0.84	0.87
16.0	0.68	0.05	0.63	0.67	0.63	0.65
32.0	0.35	0.03	0.31	0.30	0.30	0.30
64.0	0.07	0.01	0.07	0.07	0.07	0.07
128.0	0.03	0.00	0.03	0.01	0.02	0.02

VES 88

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	9.12	0.33	8.69	6.54	6.44	6.49
1.0	6.16	0.43	5.89	3.98	4.32	4.15
2.0	6.06	0.35	5.54	3.78	3.64	3.71
4.0	5.72	0.32	5.38	3.28	3.47	3.38
8.0	4.72	0.40	4.32	2.89	2.98	2.94
16.0	2.85	0.27	2.50	1.61	1.62	1.62
32.0	0.86	0.16	0.70	0.78	0.79	0.79
64.0	0.15	0.04	0.11	0.19	0.17	0.18
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 89

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	47.00	2.27	44.90	32.70	31.50	32.10
1.0	33.80	2.29	31.70	24.00	21.90	22.95
2.0	27.70	1.42	26.20	17.42	19.91	18.67
4.0	17.35	1.27	16.07	10.62	14.34	12.48
8.0	6.12	0.87	5.32	6.72	5.56	6.14
16.0	2.57	0.17	2.40	2.29	2.25	2.27
32.0	0.53	0.01	0.53	0.50	0.51	0.51
64.0	0.08	0.01	0.07	0.12	0.12	0.12
128.0	--	--	--	--	--	--

VES 90

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	4.76	0.28	4.55	3.88	3.45	3.67
1.0	2.20	0.13	2.07	1.76	1.55	1.66
2.0	1.47	0.08	1.39	0.97	0.98	0.98
4.0	1.25	0.07	1.18	0.80	0.81	0.81
8.0	1.13	0.08	1.06	0.73	0.76	0.75
16.0	0.83	0.07	0.76	0.63	0.58	0.61
32.0	0.40	0.03	0.36	0.33	0.34	0.34
64.0	0.07	0.01	0.06	0.06	0.07	0.07
128.0	0.04	0.01	0.04	0.03	0.03	0.03

VES 91

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	10.79	0.67	10.13	7.24	7.58	7.41
1.0	8.30	0.56	7.74	5.19	6.21	5.70
2.0	4.61	0.37	4.22	3.64	3.38	3.51
4.0	3.81	0.24	3.57	2.44	2.65	2.55
8.0	2.55	0.23	2.31	1.82	1.95	1.89
16.0	1.08	0.09	0.98	0.81	1.12	0.97
32.0	0.29	0.02	0.28	0.29	0.26	0.28
64.0	0.04	0.00	0.04	0.03	0.03	0.03
128.0	0.01	0.00	0.01	0.02	0.02	0.02

VES 92

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	39.80	2.99	36.80	28.50	26.80	27.65
1.0	25.90	1.90	24.10	18.48	19.37	18.93
2.0	13.73	0.92	12.81	10.25	10.97	10.61
4.0	7.17	0.34	6.83	5.47	5.01	5.24
8.0	2.52	0.23	2.37	2.64	2.64	2.64
16.0	1.73	0.06	1.65	1.12	1.21	1.17
32.0	0.38	0.02	0.35	0.35	0.37	0.36
64.0	0.09	0.01	0.07	0.05	0.07	0.06
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 93

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	24.30	0.85	23.60	16.36	14.34	15.35
1.0	9.64	1.29	8.40	7.28	8.71	7.99
2.0	6.91	0.23	6.69	5.31	4.50	4.91
4.0	4.70	0.21	4.51	3.37	3.04	3.21
8.0	1.66	0.22	1.44	1.72	1.37	1.55
16.0	1.31	0.19	1.21	0.85	1.07	0.96
32.0	0.34	0.00	0.34	0.34	0.18	0.26
64.0	0.21	0.01	0.20	0.34	0.02	0.18
128.0	0.17	0.08	0.09	0.05	0.05	0.05

VES 94

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	12.64	0.43	12.38	9.58	6.97	8.28
1.0	7.79	0.48	7.29	5.64	4.82	5.23
2.0	6.51	0.39	6.13	4.46	4.28	4.37
4.0	4.40	0.32	4.04	2.97	3.23	3.10
8.0	2.93	0.23	2.62	2.05	2.16	2.16
16.0	1.51	0.10	1.42	1.12	1.29	1.21
32.0	0.42	0.03	0.409	0.38	0.39	0.39
64.0	0.07	0.01	0.07	0.07	0.04	0.06
128.0	--	--	--	--	--	--

VES 95

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	19.55	1.14	18.43	13.28	11.85	12.57
1.0	23.20	1.90	21.30	18.97	10.33	14.65
2.0	25.60	1.59	24.00	24.30	8.15	16.23
4.0	12.88	1.17	11.77	13.95	5.89	9.92
8.0	7.21	0.32	6.90	7.11	3.24	5.18
16.0	3.38	0.09	3.12	3.27	1.53	2.40
32.0	0.28	0.01	0.26	0.49	0.42	0.46
64.0	0.12	0.00	0.12	0.16	0.27	0.22
128.0	0.07	0.02	0.06	0.06	0.03	0.05

VES 96

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	5.68	0.29	5.41	4.42	3.99	4.21
1.0	3.23	0.16	3.17	2.25	2.28	2.27
2.0	2.66	0.15	2.52	1.72	1.80	1.76
4.0	2.49	0.18	2.33	1.52	1.70	1.61
8.0	1.93	0.19	1.80	1.40	1.41	1.41
16.0	1.09	0.13	0.96	0.75	0.99	0.87
32.0	0.30	0.03	0.27	0.42	0.42	0.42
64.0	0.09	0.02	0.08	0.06	0.07	0.07
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 97

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	29.00	1.17	27.90	21.70	16.03	18.87
1.0	22.40	1.23	21.10	17.76	11.54	14.65
2.0	14.78	1.05	13.68	12.91	7.99	10.45
4.0	7.18	0.43	6.82	6.10	4.34	5.22
8.0	3.78	0.26	3.64	3.36	2.35	2.86
16.0	0.97	0.06	0.89	1.12	0.78	0.95
32.0	0.29	0.00	0.29	0.26	0.27	0.27
64.0	0.12	0.01	0.11	0.10	0.02	0.06
128.0	0.07	0.01	0.06	0.06	0.02	0.04

VES 98

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	34.40	2.54	32.00	22.90	24.10	23.50
1.0	32.30	1.64	30.60	20.30	21.30	20.80
2.0	23.90	1.96	22.00	16.19	17.27	16.73
4.0	17.27	1.29	15.73	11.73	11.78	11.76
8.0	6.42	0.57	5.96	6.21	5.44	5.83
16.0	2.14	0.16	1.99	1.86	2.00	1.93
32.0	0.52	0.01	0.51	0.40	0.40	0.40
64.0	0.08	0.01	0.07	0.03	0.04	0.04
128.0	--	--	--	--	--	--

VES 99

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	57.90	3.16	54.70	36.50	45.40	40.95
1.0	29.90	1.55	28.30	23.10	21.40	22.25
2.0	14.13	0.62	13.50	9.31	11.79	10.55
4.0	5.06	0.52	4.47	4.21	4.08	4.15
8.0	3.16	0.05	2.99	2.32	1.93	2.13
16.0	1.13	0.01	1.14	1.02	1.16	1.09
32.0	0.31	0.00	0.30	0.25	0.33	0.29
64.0	0.07	0.03	0.04	0.02	0.05	0.04
128.0	0.16	0.00	0.16	0.05	0.03	0.04

VES 100

Spacing	Resistance (Ohm)					
(m)	RA	RB	RC	RD1	RD2	RD
0.5	18.02	0.99	17.05	12.88	11.13	12.01
1.0	14.09	0.90	13.19	10.20	8.76	9.48
2.0	10.79	0.74	10.11	7.79	7.21	7.50
4.0	8.40	0.44	7.94	6.53	4.94	5.74
8.0	4.17	0.37	3.80	3.86	2.63	3.25
16.0	1.96	0.08	1.88	1.52	1.46	1.49
32.0	0.38	0.02	0.35	0.44	0.25	0.35
64.0	0.02	0.00	0.02	0.04	0.05	0.05
128.0	--	--	--	--	--	--

Appendix E2. Continued

VES 101

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	64.80	3.14	61.80	54.50	40.30	47.40
1.0	26.70	1.88	25.00	22.90	18.89	20.90
2.0	10.31	0.86	9.43	9.08	9.00	9.04
4.0	6.11	0.22	5.73	3.65	4.56	4.11
8.0	2.93	0.24	2.85	2.32	1.96	2.14
16.0	1.55	0.12	1.42	1.15	1.26	1.21
32.0	0.37	0.04	0.33	0.38	0.33	0.36
64.0	0.11	0.01	0.10	0.06	0.04	0.05
128.0	0.11	0.01	0.09	0.04	0.02	0.03

VES 102

Spacing (m)	Resistance (Ohm)					
	RA	RB	RC	RD1	RD2	RD
0.5	98.00	8.83	89.20	54.50	73.30	63.90
1.0	105.30	6.12	99.10	60.20	76.30	68.25
2.0	86.50	3.86	82.60	48.10	66.70	57.40
4.0	54.20	3.96	50.30	25.10	53.10	39.10
8.0	35.30	1.65	33.70	19.78	30.30	25.04
16.0	3.83	0.51	3.33	3.16	11.43	7.30
32.0	0.61	0.01	0.59	0.47	0.64	0.56
64.0	0.08	0.01	0.07	0.06	0.05	0.06
128.0	0.05	0.01	0.04	0.01	0.01	0.01